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A SUMMARY OF THE OROGENIC EPOCHS IN THE GEOLOGIC HISTORY OF NORTH AMERICA

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INTRODUCTION

Ever since the appearance of Dana's *Manual of Geology*, students of American geology have been familiar with the idea that the earth's crust has from time to time suffered crumpling along sinuous belts, many of which have a suggestive parallelism to the margins of the ocean basins. Of these the folding of the rocks now underlying the Appalachian Mountains serves as an example and is undoubtedly one of the most widely known. Frequent notices of such epochs of deformation appear in later works by American geologists, and in recent years the general principle has been clearly stated by Chamberlin.[1] This principle, well supported by evidence from many points of view, is that the lithosphere is not an absolutely incompetent mass, but that it has sufficient powers of resistance to the forces which tend to deform it to be able to accumulate stresses for long periods of time without perceptible deformation. These periods of accumulation are periods of quiescence, during

[1] T. C. Chamberlin, "Diastrophism the Ultimate Basis of Correlation," *Jour. Geol.*, XVII (1909), 685–93.

which base-leveling of the surface is widely extended.[1] Each period of quiescence and stress-accumulation is terminated by a short epoch of deformation during which the rocks yield to the overpowering stresses to such an extent as to produce temporary equilibrium among the stresses and thus inaugurate a new period of quiescence during which the cycle of planation of the surface may be repeated. Illustrations of the principle may be found in the history of any of the well-known mountain systems such as the Appalachians, the Rocky Mountains, or the Alps and their contemporaries.

For the purposes of the present paper, the sharp foldings and often intense plications of strata are to be discriminated from those gentler but more widespread movements generally termed "warping." With the latter are often associated faulting of the so-called normal type, and volcanic activity. These things, although different in kind, are not unrelated. There is ample evidence that rock folding is often accompanied or followed by volcanic action; and one of the phenomena attending the plication along a narrow strip may be much more widespread warping with more or less normal faulting. Nevertheless, it is not clear that all four are necessarily present as phenomena of any one period of readjustment. In the strict etymological sense, folding, warping, faulting, and volcanic action are all "oro-genic" disturbances (*oros*="mountain," *gennao*="to produce"), since each may, independently of the others, give rise to mountain forms. In this paper, however, the word "orogenic" will be used in a somewhat limited sense and applied only to those epochs characterized by prominent rock folding.[2]

Recent textbooks and other general works on the geology of North America display an interesting lack of agreement as to the number of orogenic epochs recognized, in the emphasis placed upon

[1] As an additional but quite distinct factor in lowering the lands Chamberlin also introduces the idea of a slow glacier-like creep of the continental masses outward toward the ocean basins during the quiescent periods. Since planation and crumpling may alternate, whether body creep is a real process or otherwise, such a process will not be considered in this paper.

[2] The term "revolution" is often given to these disturbances but appears to be too strong a word, in view of the frequent recurrence of such events, the world over.

them, and in the conscious discrimination between the crumplings and disturbances of other kinds. For example, Lyell[1] in 1855 makes no specific mention of such events; Dana[2] in 1875 enumerates seven more or less distinct epochs; Leconte[3] in 1891 gives four; Dana[4] again in 1895, eight more fully detailed than in 1875; Chamberlin and Salisbury[5] in 1905, six; Scott[6] in 1907, five; and Haug[7] in 1910–11, four (for North America). Dana was apparently among the first to see clearly that there had been a succession of foldings separated by periods of relative inactivity, and he early realized their importance as milestones in the progress of the earth's history. If all writers on the geology of local districts had even in recent years been as careful as Dana to distinguish between folding and mere change of level, or between folding, faulting, and volcanic activity, it would now be much easier for students of geologic history to interpret the local reports which necessarily form the basis of such general studies as this. We should not then be left in doubt as to just what an author means by the phrase, "a disturbance at the close of the Eocene" or "saecular changes in Acadia"; nor would the "Appalachian revolution" be described as extending from the close of the Mississippian to the end of the Trias and affecting most of North America.

The following list (Table I) of orogenic epochs is compiled for the use of students of geologic history without any pretense of adding materially to the existing fund of knowledge upon the subject. Like codifications in general, it may serve quite as much to call attention to the existing lack of knowledge on many points, as to present what is already known. Each of the orogenic epochs deserves to be the subject of a special investigation, for the purpose of ascertaining more exactly the date and duration of the disturbance, its relations to those in other continents, the special

[1] Sir Charles Lyell, *Elements of Geology*, 5th ed.

[2] J. D. Dana, *Manual of Geology*, 2d ed.

[3] Jos. Leconte, *Elements of Geology*, 3d ed.

[4] *Op. cit.*, 4th ed.

[5] T. C. Chamberlin and R. D. Salisbury, *Geology*, Vols. II, III.

[6] W. B. Scott, *Introduction to Geology*, 2d ed.

[7] Émile Haug, *Traité de Geologie*, Vol. II.

TABLE I

OROGENIC EPOCHS

Period	Division	Orogeny	Region Affected
Quaternary		<Santa Barbaran	California
Pliocene			
Miocene	Upper	<Antillean	Coast Ranges, Antilles, etc.
	Middle		
	Lower		
Oligocene			
Eocene	Upper		
	Middle		
	Lower		
(Paleocene)	Basal	<Laramide? <Laramide	Rocky Mountain system
Cretaceous		<Oregonian	California, Oregon, etc.
Comanchean			
Jurassic	Upper	Nevadian	Sierra Nevada, Alaska Range, etc.
	Middle		
	Lower		
Triassic		<Appalachian	Appalachians to Newfoundland
Permian			
Pennsylvanian	Upper	<Arkansan	Southern Arkansas and Oklahoma
	Middle		
	Lower		
Mississippian			
Devonian	Upper	Brunswickian	Acadia to Central Virginia
	Middle		
	Lower		
Silurian		Taconic	Gulf of St. Lawrence to the Carolinas (?)
Ordovician	Upper		
	Middle		
	Lower		

TABLE I—*Continued*

Period	Division	Orogeny	Region Affected
Cambrian		<Penokean	South of Lake Superior (with extensions)
Algonkian	Keweenawan		
	U. Huronian	<Mesabian	North of Lake Superior (with extensions)
	M. "		
	L. "		
Archean		Laurentide ? ?	(Indefinite)

phenomena characteristic of it, and the distribution of the effects in varying degree.

THE INDIVIDUAL EPOCHS OF DEFORMATION

Laurentide orogeny (*late Archean*).—This may be included in our table more for the sake of completeness than because much information can be given concerning it. In the Lake Superior district, and particularly north of Lake Superior, the Archean rocks are generally more deformed and metamorphosed than the Algonkian strata which overlie them unconformably. In nearly all other districts where the Archean has been identified the same condition prevails. Whether these facts are to be interpreted as indicating a single almost universal orogenic disturbance just before the Algonkian deposits were laid down, or several, if not indeed many, such disturbances affecting successive strips of the continent at different times, is not now determinable. Considering the periodicity and local effects of the foldings in later geologic times, the latter view is perhaps the more favored one.

Mesabian orogeny (*late Middle Huronian*).—In northeastern Minnesota and adjacent portions of Canada the Upper Huronian (Animikean) strata lie gently tilted upon the contorted and partly metamorphosed beds of early Huronian and Archean age. The interval during which these folds were made and truncated has been termed the "Eparchean interval" by Lawson,[1] and by him

[1] A. C. Lawson, *Bull. Dept. of Geol. Univ. of Cal.*, III (1902), 51–62.

pronounced the most important stratigraphic break in the pre-Cambrian sequence. The emphasis thus placed on the mid-Huronian unconformity is due to the consideration of only a limited belt north of Lake Superior. When a much larger region is examined, this glorification of the mid-Huronian orogeny appears unwarranted. The nearly parallel relations of all the Huronian series south of Lake Superior proves that there the rocks were not folded in the mid-Algonkian. Recently Leith[1] has reviewed our knowledge of the mid-Huronian unconformity, showing that a conspicuous angular discordance between sedimentary beds of pre-Cambrian age has been found farther east in the Sudbury and Cobalt districts of Ontario, and on into Quebec. It is not certain that it is the same unconformity and that it is of mid-Huronian age in all of these localities, but it seems within the bounds of probability that such a correlation may be established in the future.

The lack of fossiliferous strata of the ages involved still prevents us from correlating the Algonkian sections of widely separated parts of the continent. Were it not for this fact, we might be able to ascertain in large measure the real extent of the Mesabian folding. Suggestively similar relations are found in the Rocky Mountains from Wyoming to Arizona, for in those states comparatively unaltered late Algonkian rocks rest upon highly contorted and metamorphosed sedimentary beds usually referred to the early Algonkian. The outcrops of these folds are largely covered by later strata and in isolated exposures the structural lines do not seem to follow a single general direction.

Penokean orogeny (*late Algonkian*=(?) *post-Keweenawan*).—Again the Lake Superior region supplies the standard of reference. After the extrusion of the Keweenawan basalts overlying the essentially undeformed Upper Huronian strata, the region immediately south of Lake Superior was compressed into folds trending nearly east and west. This apparently produced the Lake Superior synclinorium. In parts of northern Michigan the folding was intense, leaving the strata quite schistose and in nearly vertical attitude. Large batholitic intrusions accompanied the folding. On the north side of Lake Superior, however, the deformation was

[1] *Proceedings of the Internat. Geol. Congress*, Toronto, 1913.

so mild that on the Mesabi Range and in Thunder Bay the Upper Huronian beds are still only gently inclined. Westward the Penokean orogenic disturbance can be traced with some confidence to the Cuyuna Range of central Minnesota, and southward perhaps as far as the Baraboo district of southern Wisconsin, where rocks usually referred to Middle or Upper Huronian are highly folded. Eastward it extended an unknown distance into Ontario.

In reference to the late Algonkian orogeny, as to its predecessors, it is not yet possible to trace the extension of the folding much beyond the Lake Superior district because of the present uncertainty as to the ages of the strata. It is worth noting, however, that rocks believed to be of late Algonkian age, more or less strongly folded before the deposition of the Paleozoic beds, exist in the Black Hills of South Dakota, in Colorado, Montana, Utah, Arizona, and perhaps other western states and Canadian provinces. These beds are now so generally covered by younger rocks, that no continuous system of mountain folds has been worked out. The two Algonkian epochs of folding in the Rocky Mountains have been discussed by S. F. Emmons.[1]

Taconic orogeny (late Ordovician).—Because its effects are most evident in New England, the "Taconic revolution" was recognized more than fifty years ago by that group of geologists among whom Dana was ever a leader. Even today it would be difficult to improve upon the description, given in the early editions of Dana's *Manual of Geology*, of this signal event in the history of eastern United States. Because of its relation to fossiliferous strata, it is the first of the deformational epochs of which the age can be determined with a high degree of accuracy in the geologic scale and traced with confidence over a large area. Its effects in closely folded and even metamorphosed strata have been traced from the vicinity of Quebec (city) in Canada south through Vermont and western Massachusetts to New York City. Strong suggestions that it extended southwestward as far as Virginia have already been cited by Dana and others, but it is doubtful whether we shall soon know how far in that direction the crumpling was felt. The

[1] S. F. Emmons, "Orographic Movements in the Rocky Mountains," *Bull. Geol. Soc. Am.*, I (1890), 245–86.

Cambro-Ordovician rocks were folded in the Gaspé Peninsula and as far as northwestern Newfoundland. The belt of deformation was apparently wide enough to include most of New Brunswick, perhaps Nova Scotia, nearly all of Maine, and probably even Rhode Island. The warpings and changes of level which accompanied the folding affected a large part of the eastern interior of the United States and doubtless Canada. These movements have lately been summarized by Schuchert.[1] In western United States, however, there appears to be evidence of nothing more than a change of relative altitude between the sea-level and the land surface.

Brunswickian (late-middle Devonian).—Although the folding of the older rocks in eastern Canada, just before the deposition of the Carboniferous measures, was well known to Logan, Dawson, and other pioneer Canadian geologists, and is noted by Dana in all the editions of his *Manual*, the event has received much less attention than it deserves from geologists in general. Several of the more recent textbooks of geology make no mention of the occurrence, and it is sometimes ignored in discussions of Devonian paleogeography where it is a factor of importance. Dawson's description,[2] although reflecting some provincial bias, may serve to call attention once more to the importance of the Devonian disturbance. He says in part:

> The whole surface of Acadia was thrown into a series of abrupt folds—great masses of plastic granitic matter invading every opening in the shattered masses. This period surpasses every other, in the geological history of the eastern slope of the American continent, in its evidence of fracture of the earth's crust. To this period we must refer the greater part of the intrusive granites of Eastern America, and to it also is referable the greater part of the metamorphism of the Silurian rocks, and the origin of the numerous metallic veins by which these are traversed.

As there appears to be in current use no name for this folding, the term "Brunswickian" is suggested to meet the evident need.

[1] Charles Schuchert, "Paleogeography of North America," *Bull. Geol. Soc. Am.*, XX (1908), 488–89.

[2] Sir William Dawson, *Geology of Nova Scotia, New Brunswick, and Prince Edward Island*, 2d ed. (1891), pp. 665–66.

The tract within which the rocks were crumpled during this epoch stretches from western Newfoundland through Cape Breton Island, Nova Scotia, New Brunswick, Maine, and probably southern New England, and was wide enough to include on the northwest the Gaspé Peninsula in eastern Quebec. Farther southwestward, the evidence becomes scanty, but the marked unconformity beneath the basal Carboniferous beds near Boston and in Rhode Island suggests that it extended at least as far as Long Island Sound. Willis[1] has shown that the great thickness of late Devonian and early Carboniferous clastic rocks in Maryland and Virginia strongly suggests the rise of mountain ranges immediately to the east, in the region now occupied by New Jersey and eastern Virginia. Using the same criterion, the dwindling and eventual disappearance of such formations between Virginia and Alabama suggest that the folding did not extend the entire length of the present Appalachian Mountain belt.

Although there are some discrepancies in the evidence and hence some disagreement among writers on the subject, the testimony of unconformities on the one hand, and of thick clastic formations on the other, indicates that the Brunswickian disturbance culminated after the middle of the Devonian but considerably before the close. In Acadia the folds had been truncated before the Lower Carboniferous (Mississippian) period and beds of that age were deposited upon the eroded stubs.

Arkansan (*mid-Pennsylvanian*).—The folded structures underlying the mountains of Arkansas and Oklahoma were made, as nearly as can be inferred from current correlations, in the latter part of the Pennsylvanian period. Thus in the central Arkansas coal field the deformation followed the laying-down of the lower Pennsylvanian coal measures,[2] but no younger strata exist there. In the Arbuckle Mountains of Oklahoma it occurred after the deposition of the Caney shale (early Pennsylvanian?) and before that of the Franks conglomerate (late Pennsylvanian). In the Wichita Mountains still farther west in the same state, the folds had been truncated before the deposition of the Oklahomian (early

[1] Bailey Willis, *Geol. of Maryland*, IV (1902), 23–93.

[2] A. J. Collier, "Coal Fields of Arkansas," *U.S. Geol. Survey, Bull. 326*, p. 24.

Permian) red beds. Thus it is the conclusion of Taff[1] "that the Arbuckle uplift [= crumpling] began near the middle and culminated near the close of Pennsylvanian time, previous to the deposition of the red beds."

The folds are sharply compressed and cut by overthrusts and other faults as in the Appalachians, but there are no contemporaneous igneous intrusions and the rocks were not much metamorphosed. In Oklahoma, the Arkansan system of folds trends east by south, but the strike becomes more nearly east and west in Arkansas. Extensions of the system in the line of strike are unknown, because both to the west and to the east the folded beds are concealed by overlapping horizontal strata of later age. The same is true to the south. Southwestward in the Llano[2] district of central Texas, early Pennsylvanian strata are the youngest involved in the very mild folding and faulting which the stratified rocks now show. Their deformation may perhaps be related to the Arkansan disturbance. North of the Arkansas River the Carboniferous strata still lie almost horizontal, and in Kansas there is not so much as a disconformity between the early and later Pennsylvanian strata. In the Ozark region of Missouri and northern Arkansas the very gentle flexures and normal faults may possibly be referable to this epoch, but as yet the evidence is of little weight. Slight deformation of the Mississippian and older strata along the Mississippi River in Illinois, Iowa, and Missouri, and the making of the LaSalle anticline of Illinois seem to have in large measure preceded the Pennsylvanian, and if so, are not to be correlated with the Arkansan orogeny. The disturbance seems too slight to include, in the present list, as a distinct orogeny.

The folding of the Ouachita beds has been referred by Dana and others to the Appalachian revolution. However, unless published correlations are seriously in error we must conclude that the Ouachita folds had been formed and truncated before the deposition of the latest Pennsylvanian sediments, whereas the Appalachian folds were not begun until after the early Permian strata had been

[1] J. A. Taff, "Geology of the Arbuckle and Wichita Mountains," *U.S. Geol. Survey, Prof. Paper 31*, p. 80.

[2] Sidney Paige, *U.S. Geol. Survey, Llano Folio, No. 183*, Texas.

laid down. It is now generally agreed that the climax of that disturbance came near the close of the Permian period. If these correlations are correct, we must then recognize two separate orogenic epochs. There is apparently ground for correlating the Arkansan crumpling with that which produced the Armorican and Variscian systems of western Europe, which Haug[1] assigns to the opening of the Stephanian (upper Pennsylvanian) epoch.

Appalachian orogeny (*late Permian*).—This epoch of folding was the first to be well recognized in America and is without doubt the one now best known to students of American geology. Its effects have been traced from Alabama to New York and from southern New England to Newfoundland. The southeast limit of the area affected is everywhere concealed either by the ocean or by much younger undeformed deposits. To the northwest the limit is rather sharply defined where the folds give way to the horizontal strata of the Cumberland and Alleghany plateaus. There are, it is true, some broad gentle swells in these apparently horizontal beds, and those flexures were probably produced at the same time as the Appalachian folds. In western New England and adjacent parts of New York and Canada, the limits are less easily ascertained because little now remains but pre-Devonian or even pre-Silurian strata, and those had been already folded either in the Brunswickian or the Taconic epoch or both. Available evidence, although meager, indicates that the crumpling and overthrusting at the close of the Permian were even more intense in the Piedmont and New England regions than in the Appalachian Mountains themselves.

Reasons have been given above for believing that the Ouachita region was not deformed at this time. If this conclusion is correct it may be said that the western three-quarters of North America was not affected by the Appalachian orogeny except that there were, over large areas, mild epeirogenic movements which made notable changes in the relations of land-masses and seas.

Nevadian (*late Jurassic*).—The folding of the rocks in the Sierra Nevada at the close of the Jurassic was established by Whitney and his associates about half a century ago, but it seems to be less widely understood that the entire west coast of North,

[1] Émile Haug, *Traité de Geologie*, II, Part I (1910), p. 829.

and probably South, America was affected at about the same time. Smith[1] has called this the "Cordilleran revolution," but since it affected less than half of the great mountain tract of the West, to which we usually apply the name "Cordilleras," the designation seems likely to mislead and will perhaps tend to perpetuate a widespread popular misconception of the western mountains as all belonging to one great system instead of two or three systems. The term "Nevadian" has therefore been substituted in this paper, in allusion to the Sierra Nevada.

The system of folds made at this time trends in a north-northwesterly direction more or less parallel to the Pacific shore of today, but by no means coinciding with it. South of the typical area in the Sierra Nevada, the Jurassic folds may be traced into western Mexico and probably Lower California. Northward along the strike it may be traced through Oregon, the Cascade and Olympic ranges of Washington, the Coast Range of British Columbia, and the St. Elias and Alaskan ranges beyond. There are even suggestions of its extension into the Aleutian chain. Its western limit is concealed everywhere by the Pacific Ocean or by unconformably overlapping younger deposits which have themselves been folded subsequently. To the east and northeast, the Nevadian system of folds is involved with the overlapping Laramide system and it therefore becomes a difficult task to discriminate the two. The eastern limit of the Nevadian folds seems to be reached approximately near Bisbee, Arizona, where Comanchean sediments overlie unconformably the deformed Paleozoic strata, and the ill-defined boundary may be traced northwestward through eastern Nevada,[2] western Idaho, central British Columbia, the upper Yukon valley in Canada, and central Alaska in general. There is good evidence that it did not affect Alaska north of the Yukon nor the eastern ranges of Mexico.

[1] J. Perrin Smith, *Science*, XXX, 346–51.

[2] It should be noted that much of this region has been so modified by igneous intrusions and faulting of much younger date that the results of the Nevadian orogeny have been in large measure lost to view. It is these later disturbances which have made the present Basin Range province of Nevada distinct from the Sierra Nevada province of California, though both were once parts of the Jurassic belt of folding.

Throughout this great belt, which doubtless has extensions in Asia on the west and also in South America, the folding was accompanied or followed by the intrusion of enormous masses of granitic rocks, the commonest types among which are granodiorite and quartz-diorite. In the intensity of the folding there was much variation, but through California and Washington, at least, it was great.

Although nearly all writers on the subject agree in assigning this epoch of diastrophism to the Jurassic period or its close, there is considerable difference of opinion as to a more precise date. The Alaskan geologists have usually referred it to the middle Jurassic. Stanton and the earlier Californian geologists assigned it to the end of the Jurassic. According to J. P. Smith it took place just before the Portlandian epoch of the late Jurassic. These differences of opinion are probably in large part due to the difficulty of correlating the faunas of such widely separated countries as California, Alaska, and Europe, rather than to any real difference in the date of the disturbances in the several regions.

Oregonian orogeny (*post-Comanchean*).—In the literature of Pacific coast geology there is much evidence to indicate that the deposition of the Shasta and related series (Comanchean) was followed by crumpling of the strata, accompanied by the intrusion of igneous rocks, especially peridotitic varieties, now changed to serpentine. The phenomena have been found interruptedly from southern Mexico through California, Oregon, and Washington to southeastern Alaska. In all cases they lie near the coast, and in no instance are they known to occur east of the area affected by the Nevadian orogeny. As in that preceding orogenic epoch, so here, the folds appear to be roughly although not exactly parallel to the edge of the Pacific basin. In most districts the folding was of moderate intensity but in southwestern Oregon the Comanchean strata stand on edge in isoclinal attitude.

All inferences as to a period of deformation between the lower and the upper Cretaceous necessarily rest upon the correlations of the strata involved. On the whole, the Comanchean and Cretaceous beds of the Pacific province are only sparingly fossiliferous and many of the species, such as *Aucella*, are of large vertical range

and hence of doubtful value in correlation. The various writers on Pacific coast geology are, therefore, by no means in harmony as to the ages of some of the strata. The existence of a conspicuous angular unconformity is conceded. Above this conformity lies either the Chico (upper Cretaceous) or certain Eocene formations. Below it lie much more highly folded and locally even metamorphosed rocks which may in general be referred to as the *Aucella* beds. These have yielded both invertebrates and plant remains. The former have been classified usually as lower Cretaceous, but sometimes as late Jurassic; the plants nearly always as Jurassic. At present the balance of qualified opinion seems to be in favor of regarding the system as largely Comanchean but in part late Jurassic. This orogeny, if it was a fact, is among the least-known in the later part of the earth's history.

Laramide orogeny (*post-Cretaceous*).—The deformative movement which gave birth to the Rocky Mountain system is one of the three best-known events of its kind in North America. The folds thus produced trend north by west from southeastern Mexico to western Texas, thence across New Mexico, Colorado, Wyoming, Montana, western Alberta, and the northwest provinces of Canada into northern Alaska, finally reaching the Arctic Ocean at Cape Lisburne. Although there may have been folding in Central America also at this time, the fact that the same region suffered a further crumpling in the midst of the Tertiary has largely if not entirely obliterated evidences of earlier deformation there. The post-Cretaceous folds reappear, however, in full force in the Andean system of South America. On the eastern or inland side of the system the limits are generally distinct, for the folds pass somewhat abruptly into the nearly horizontal strata of the Great Plains. There are, however, some outlying tracts of disturbance characterized by gently tilted strata and laccolithic intrusions, as in the Black Hills of South Dakota and the Uvalde district of Texas. The same general condition seems to prevail throughout the extent of the system. On the westward or Pacific side the limits are much less definite because the post-Cretaceous folds are not easily distinguished from those of the Nevadian epoch, except where late Cretaceous strata are present; and it is an embarrassing fact that outcrops of

rocks of this age are comparatively scarce in the plateau region between the Pacific and Rocky Mountain systems. Such facts as are now available suggest that the western limit of the Laramide disturbance passes through central Arizona, eastern Nevada, western Idaho, and central British Columbia to the Yukon valley. This tentative division of the provinces is supported by the fact that the Comanchean rocks in the Bisbee region, Arizona, have been gently tilted and faulted but only slightly folded; the Cretaceous rocks near the headwaters of the Yukon River have been folded, and the same is true of beds of similar age on the northeast side of the Seward Peninsula in Alaska. On the other hand, along the Pacific slope in California, Oregon, Washington, British Columbia, and Alaska, wherever both upper Cretaceous and Eocene strata are present, they are generally either conformable, or separated by a mere disconformity. At the north end of the Sacramento Valley, California, the upper Cretaceous strata still remain almost horizontal.

There is substantial agreement among geologists that the Laramide orogeny marks approximately the close of the Cretaceous period. Nevertheless, until the existing controversy over the correlation of the latest Cretaceous and earliest Eocene formations of western United States is satisfactorily settled, the exact date of the deformation remains uncertain. In some parts of western Wyoming and Colorado there are suggestions of two episodes of crumpling separated by a short but definite epoch of erosion and the deposition of sediments in the earliest Eocene (Paleocene). Similar phenomena are reported from Yellowstone Park and southwestern Montana.

Throughout its range the Laramide orogeny was marked by great volcanic activity, which manifested itself in the production of laccoliths and volcanic cones with their associated dikes, sills, flows, and fragmental deposits. Locally in parts of Idaho and western Montana granitic batholiths were intruded at this time, but they are much less characteristic of the Laramide system than of the Nevadian.

Antillean orogeny (*middle-to-late Miocene*).—It has long been known that the dominant structural features of the California and Oregon coast ranges were produced in the Miocene period. To

the north the deformation may be traced through the Oregon Coast Range to the Olympic Mountains of Washington, thence through the coastal islands of British Columbia, to the southern Alaska coast, and on into the Aleutian chain. To the south the same deformation is reflected in the tilted and more or less disturbed Eocene beds of Lower California. Beginning in southwestern Mexico, a belt of complexly folded and faulted rocks with volcanoes extends through Honduras and other Central American states, through the West Indian islands to Venezuela. These folds involve Eocene, Oligocene, and even Miocene strata, but are unconformably overlapped by the undisturbed Pliocene. Eastward and northward the effects of the Antillean disturbance wane into gentle folds and finally into mere warping, accompanied by normal faults and volcanic structures. Decided folding of early Tertiary strata extends as far east as southwestern Nevada, the Blue Mountains of Oregon, the Cascade Range and Okanagan regions of Washington, at least the western part of the Fraser River basin of British Columbia, and the upper Yukon valley. In Alaska it affected the Kuskokwim Mountains and Nunivak Island, but not the Mt. Wrangell district or the region north of the Yukon River. Even as far east as the Rocky Mountain states of Wyoming, Colorado, and Montana, the Eocene and Oligocene strata are locally somewhat tilted and broken by normal faults, suggesting that the Antillean disturbance made itself felt in a mild way as far east as the edge of the Great Plains. Eastern United States suffered no folding but there was some warping, especially in Florida and the Gulf states.

By comparison with the Nevadian and Laramide disturbances, it will be seen that the Antillean orogeny affected an area as large as, or even larger than, the Nevadian, and again the strongest folding is found nearest the Pacific coast. Almost throughout the range of the Miocene disturbance, whether mild or severe, volcanoes broke forth and scattered their lavas and fragmental deposits far and wide. In that respect the epoch may have exceeded even the Laramide orogeny. Batholiths of the Antillean epoch are reported from the Cascade Mountains of Washington and British Columbia; they are probably the youngest known.

Nearly all authorities agree in assigning this event to the middle Tertiary, and by many it is referred to the close of the Miocene period. In the John Day basin of central Oregon, according to Merriam and Sinclair,[1] gentle folding closed the deposition of the early Miocene sediments, and preceded the Columbia lava outflows. In the northern part of the Cascade Mountains of Washington a similar event is recorded by Smith and Calkins.[2] The most accurate estimate of the date of the deformation seems to be that which has been made in the Coast Range of southern California by Arnold[3] and others, who show that it followed the deposition of the lower-middle Miocene sediments (Monterey shale), and preceded that of the upper Miocene (Santa Margarita).

Santa Barbaran orogeny (early Pleistocene).—Recent studies in the oil-bearing districts of southern California have brought to light the fact that, although the region was considerably affected by the Antillean orogeny, it was as much or even more closely compressed after the deposition of the marine Pliocene formations. Thus in the region about Santa Barbara, the Pliocene has been not only folded but overturned, and locally lies beneath overthrust masses of earlier Tertiary strata. In the Coalinga oil district[4] on the southwest side of the great valley of California, the effects of the Santa Barbaran disturbance are much more pronounced than those of the middle Miocene folding.

Although it has long been known that warping, gentle changes of level, and normal faulting proceeded on a large scale early in the Pleistocene period, particularly in western United States, there are but few places outside of California where actual folding has been assigned to this age. Possibly we should refer to the Santa Barbaran folding the gentle flexures in the late Miocene lavas of central Washington and the Cascade Mountains, as well as the latest gentle folds in the Olympic Range of the same state. To the

[1] J. C. Merriam and W. J. Sinclair, *Bull. Dept. of Geol. Univ. of Cal.*, V (1907), 174.

[2] G. O. Smith and F. C. Calkins, *U.S. Geol. Survey, Snoqualmie, Wash., Folio, No. 139.*

[3] Ralph Arnold, "Environment of the Tertiary Faunas of the Pacific Coast of the United States," *Jour. Geol.*, XVII (1909), 509–33.

[4] *Ibid.*, "Geology of the Summerland Oil District, California," *U.S. Geol. Survey, Bull. 321.*

north the comparative scarcity of Pliocene strata renders the tracing of the folding in that direction more difficult. In the region of the Antilles, Pliocene strata are much more abundant, but there they are apparently not folded and only very gently tilted.

GENERAL CONSIDERATIONS

The table of orogenic epochs given above is still far from complete, and tabulations of this kind are necessarily subject to correction. The most evident defects relate to the pre-Cambrian. Until the pre-Cambrian terranes in widely separated districts have been correlated much more securely than at present, we shall not know how many orogenic epochs fell within the Archeozoic or even the Proterozoic era; and we shall know but little of the true extent of those ancient orogenies which are already recognized as facts. Again, the effects of some orogenic movements, especially near the borders of the continent, have been so masked by later crumplings or have been so largely buried beneath younger sediments or by the sea itself, that the record, once clear, is now largely destroyed. Thus we know nothing of the southwestward extension of the Appalachian system of folds beneath the sediments of the Gulf coastal plain; and on the Pacific slope the complex structure and metamorphism render it difficult to get information about any orogenic epochs which that region may have suffered before the Jurassic. Some of these limitations will gradually be removed as the making of critical local studies progresses.

In most of the known epochs of orogeny, a single elongate belt was affected, while much broader surrounding regions experienced nothing more revolutionary than changes of level, gentle warping, a little faulting, and sometimes volcanic activity. Most of the folded belts were thousands of miles in length and a few of the best-known measured many thousands. The Laramide system of the Americas extended over about half of the earth's circumference, and the Miocene folded belts are long enough to reach more than one and a third times around the globe.

The belts vary from 50 to more than 600 miles in width and in Asia locally exceed 1,000 miles. The average width of the crumpled zones is, however, between 100 and 300 miles. Where the folding

is closest, the zone is likely to be narrower than where the folds are open and wavelike. If this is generally true, it may mean that the narrowest places are narrow because they have been the most compressed.

The crumpled strips produced at each epoch overlap more or less, but rarely coincide. Thus the Mesabian orogeny affected the north side of Lake Superior while the Penokean affected the south side. Again, the Laramide and Nevadian zones are largely distinct, although they overlap considerably. Successive foldings in the same general region were usually roughly parallel to each other and resulted in an ever-widening belt of crumpled strata. In North America we have at least three such composite zones, the Cordilleras in the west, the Atlantic system in the east, and the ancient systems of the Canadian shield where nearly all structures trend east by north.

Although overlapping of the limits is common, it is rare that the locus of greatest intensity occurred twice in the same place. The Sierra Nevada region, intensely folded in the Jurassic, was but little affected in the Miocene, although the Coast Range to the west and parts of Nevada to the east both suffered folding at that time. Intense crumpling produces a thickening of the deformed strip and therefore increases its competency to resist further thrusts and to transmit them to regions beyond. It seems probable, also, that the intrusion of great batholiths such as accompanied the Nevadian orogeny must notably strengthen the mass within which they congeal, and thus render it less liable to yield to compression in later periods.

Many students of the subject have shown that some of the disturbances were inter-continental. For example, there appears to have been some crumpling in northwestern Europe, corresponding approximately in time to the Taconic orogeny in eastern North America. The Brunswickian likewise has its counterpart in the mid-Devonian deformation in Scotland. The Arkansan seems to correspond in time to the Armorican orogeny of western Europe, the Nevadian to the folding of the eastern interior ranges of Asia, and the Antillean almost unquestionably to the rearing of the great Alpine system of Eurasia. These striking correspondences strongly

suggest that the orogenic forces are not local but planetary. Chamberlin[1] regards the crumplings on two sides of an ocean as a result of the subsidence of the earth segment beneath that ocean.

The folding along the relatively narrow belts was nearly always accompanied by more or less warping and volcanic activity over adjacent, much wider areas. The warpings have not infrequently disturbed stream activities and strand-lines over large portions of continents, as at the close of the Permian and in the Miocene. The transitions from the folded to the merely warped areas are in some cases abrupt and in others gradual, through gently folded or faulted tracts. Where the demarcation is sharp, as the east side of the Rocky Mountains, there may have been some original line of weakness such as a decisive initial dip; whereas in the cases of gradual transition the underlying mass may be tolerably uniform in resistance, and may therefore permit the force to be transmitted through it and die away gradually.

In several instances observed relations indicate that the most effective compression was accomplished on the side toward the ocean basins. Thus we have the closely appressed folds and overthrusts of the Laramide system in western Montana and Idaho changing into gentle open folds farther east. Again, in the Appalachian orogeny, the Rhode Island district was more closely crumpled and metamorphosed than Pennsylvania on the northwest side of the deformed zone; and a like comparison may be made between the region of great overthrusts and isoclinal folds in North Carolina on the southeast and the open arches of northern Alabama on the northwest. For at least some of the orogenic systems there seems to be a significant arrangement of the batholithic intrusions on the seaward side of the zone, with superficial volcanoes and laccoliths along the landward side. This is perhaps only another way of saying that the batholiths are characteristic of the more intensely folded parts of the zone, while the superficial volcanic features occur where the rocks have been less deformed. It is admitted, however, that this distribution of intensities and phases of igneous activity cannot be demonstrated for many of the epochs and may be accidental rather than significant.

[1] T. C. Chamberlin and R. D. Salisbury, *Geology*, I (1904).

There is a conception familiar to present-day geologists, that the great mountain systems coincide with geosynclinal depressions in which sediments have previously accumulated to unusual thicknesses. The Appalachians have served as a classic example. In the opinion of some, the trough was deepened by lateral pressure and the deposition of the sediments thus invited. The extreme isostasists, on the other hand, view the trough as a result of the loading of the strip by the deposition of thick sedimentary beds. Willis urged the importance of the initial dips developed in these geosynclines in predetermining the locus of the folded belt when the compressive force became effective. It is prudent, however, not to assume that there is a causal relationship between belts of thick sediments and subsequent mountain folding, merely because one preceded the other. Much of the thickness of the Appalachian sediments was directly due to the ruggedness of the land-mass of Appalachia, from time to time in the Paleozoic era; and it should be recalled that at the close of the Permian not only the Appalachian geosyncline, but a still broader region to the east, was intensely deformed. The observed relations may be stated in another way, namely, that sediments accumulate rapidly along mountainous coasts, and that coasts in turn are liable to repeated crumplings, for reasons not here discussed. Hence the two phenomena are generally associated.

In each great portion of the continent there have been successive roughly parallel crumplings, separated by long periods of quiescence, and, in some cases at least, eventually terminated by a cessation of activity which has endured down to present times without premonition of change. In the Lake Superior region, three, and probably many more orogenies, were passed before the Cambrian, but in all subsequent time there have been none. In the Atlantic mountain system three or more crumplings before the Triassic have been followed by a stability prolonged until the present, and with no suggestions of an end. In the western Cordilleras the orogenic activities seem to have slumbered until the Jurassic, but since then the region has been subjected to repeated compressions and is the one in which future disturbances are most likely to take place.

When the earlier and later orogenic epochs are compared, there appears a suggestive sameness running through the whole series. Although we are still unable to express geologic chronology in terms of years, an estimate expressed in terms of epochs or thickness of strata suggests that the crumplings were no more frequent in the Paleozoic era than in later periods; nor do the facts seem to imply that the earlier disturbances were more intense or more widespread. In fact, if we should take the available data without making allowance for the greater loss of record pertaining to the earlier periods, we might reach the improbable conclusion that they had been more widespread, more severe, and more numerous since the Triassic than before.

THE CAUSES OF OROGENIC EPOCHS

It is beyond the purpose of this paper to discuss the origin of earth folding.

From the very nature of its effects the cause of orogenic epochs must be sought in tangential compression affecting the rocks of the earth's outer shell. As to the origin of that force there is, however, no unanimity. It constitutes a problem for which several hypotheses have been offered. These are explained and critically analyzed in the writings of Dana,[1] Reade,[2] Chamberlin,[3] Willis,[4] Taylor,[5] Leith,[6] and others.

[1] J. D. Dana, *Manual of Geology*, 4th ed. (1895).

[2] T. Mellard Reade, *Origin of Mountain Ra*[illegible]5.

[3] T. C. Chamberlin and R. D. Salisbury, *Geology*, I (1905).

[4] Bailey Willis, "Research in China," *Carnegie Institution*, II (1907).

[5] F. B. Taylor, "Origin of the Earth's Plan," *Bull. Geol. Soc. Am.*, XXI (1910).

[6] C. K. Leith, *Structural Geology* (1913).

THE STRENGTH OF THE EARTH'S CRUST

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PART VI. RELATIONS OF ISOSTATIC MOVEMENTS TO A SPHERE OF WEAKNESS—THE ASTHENOSPHERE.[1]

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INTRODUCTION AND SUMMARY

In studies on the nature of isostasy it is necessary to distinguish between, first, the existence of isostasy; second, the limits of isostatic equilibrium; and third, the mode of maintenance of this equilibrium.

The first has long been known, the knowledge of the existence of some relation of density counterbalancing elevation having been gradually developed since the middle of the nineteenth century through the determination of the local deviations of the vertical as shown by the comparison of the astronomic and geodetic latitudes for the same station. This was a problem which arose in both astronomy and geodesy. It was found, when the attractive effect of the mountain regions was computed, that they did not deflect the vertical at adjacent stations as much as was to be expected from their visible masses. The phenomenon was first pointed out

[1] An abstract of Parts V, VI, and VII of this series was given at the April, 1914, meeting of the American Philosophical Society at Philadelphia under the title, "Relations of Isostasy to a Zone of Weakness—the Asthenosphere." See *Science*, XXXIX, 842.

by Petit in 1849.[1] Archdeacon Pratt of Calcutta showed a few years later that whereas a discrepancy of 5.2″ existed between the geodetic and astronomic latitudes of Kalianpur and Kaliana, the calculation of the effect of the Himalayas called for a difference of 15.9″.[2]

These facts were definitely formulated into a theory of isostasy by the Astronomer Royal of Great Britain, G. B. Airy, within a year following the appearance of Pratt's paper,[3] though it remained for Dutton to recognize the large geologic significance and to coin for the relations of elevation and density the word isostasy.[4] Following this Putnam and Gilbert showed by gravity measurements that a considerable degree of regional isostasy existed over the United States.[5] Since then has appeared the much more detailed work of Hayford and Bowie, the computations made by the computing office of the United States Coast and Geodetic Survey under their directions making possible this present investigation.

Thus there has developed through more than half a century evidence beyond controversy which shows that the earth's crust in its larger relief and, within certain limits, even its smaller features, such as the great plateaus and basins, rests more or less approximately in flotational equilibrium.

The second division of the larger problem of isostasy, that of the areal limits and degree of perfection of isostatic adjustment, is the subject which has been dealt with in the previous parts of this investigation. It has been found that, although the relations of continents and ocean basins show with respect to each other a high

[1] "Sur la latitude de l'Observatoire de Toulouse, la densité moyenne de la Chaine des Pyrénées, et la probabilité qu'il existe un vide sons cette chaine," *Comptes rendus de l'Acad. des Sc.*, XXIX (1849), 730.

[2] "On the Attraction of the Himalaya Mountains and of the Elevated Regions beyond Them, upon the Plumbline in India," *Phil. Trans. Roy. Soc.*, Vol. CXLV (1855).

[3] G. B. Airy, "On the Computation of the Effect of the Attraction of Mountain Masses as Disturbing the Apparent Astronomical Latitude of Stations in Geodetic Surveys," *Phil. Trans. Roy. Soc.*, Vol. CXLV (1855).

[4] "On Some of the Greater Problems of Physical Geology," *Bull. Phil. Soc. Wash.*, XI (1889), 53.

[5] *Bull. Phil. Soc. Wash.*, XIII (1895), 31–75.

degree of isostasy, there is but little such adjustment within areas 200 to 300 km. in diameter, or of limited differential relief. Individual mountains and mountain ranges may stand by virtue of the rigidity of the crust. Even under the level plains equally great loads are permanently borne, loads produced by widespread irregularities of density not in accord with the topography above. Isostasy, then, is nearly perfect, or is very imperfect, or even nonexistent, according to the size and relief of the area considered.

The third division, the mode of maintenance of isostasy and its bearings on problems of the crust, remains to be considered. This condition of isostatic equilibrium exists at present in spite of the leveling surface actions and compressive crustal movements of all past geologic time. There must be, consequently, some internal mode of restoring more or less perfectly an isostatic condition, either by frequent small movements, or by more infrequent and larger ones.

Erosion and sedimentation result in a lateral transfer of matter, and to maintain isostasy there must be some lateral countermovement in the earth below, but in regard to how or where or when this is done, and as to what are its effects, there has been no unanimity of opinion, nor convincing demonstration.

In considering the problems of crustal dynamics some authors have regarded earth shrinkage and consequent tangentially compressive forces as controlling the nature of diastrophism, including movements of both orogenic and epeirogenic character; others, the advocates of extreme isostasy, have thought to see even in folding only the secondary effects of movements maintaining isostatic equilibrium. The first point of view emphasizes the strength and elasticity of the crust, with long-deferred periodic discharge of stress. The second point of view calls for an interpretation based on the weakness and plasticity of the crust, with resulting nearly continuous small movements restoring the delicate vertical balance destroyed by gradational actions. To what degree are the two points of view compatible and within what limits is each dominant? The problem of this chapter involves, therefore, not only the mode but the limits and effects of the movements which more or less completely maintain or restore isostasy.

The method of attack is largely one of exclusion. By showing what hypotheses cannot apply, the way is prepared for conclusions in better accord with the fields of fact and theory.

The results show that conditions of isostatic equilibrium cause the light and high segments to press heavily against the adjacent lower and heavier ones, most heavily above. The tendency is consequently for the high areas to spread with a glacier-like flow over the low areas. This tendency, however, is effectively resisted by the strength of the crust. Upon the disturbance of equilibrium by erosion and deposition there are two kinds of stresses produced which tend to restore equilibrium. The first is a tendency of the heavy column to underthrust the lighter, but it could never produce compression and folding at the surface. This force would be most effective under the hypothesis of great crustal weakness, so that the vertical stresses could be transmitted in a horizontal direction within the lithosphere as in a fluid. Even in that case, however, it would not be the dominating force. The actual isostatic movements consist of a rising of the eroded areas, a sinking of those which are loaded. This involves shear or flexure around their boundaries. The columns must be large enough so that the excess or deficiency of mass can become effective in producing deformation. When the accumulating vertical stresses have overcome the strength of the crust, the excess pressure from the heavy area is transmitted to the zone below the level of compensation. This deep zone is in turn the hydraulic agent which converts the gravity of the excess of matter in the heavy column into a force acting upward against the lighter column and thus deforms the crust of the eroded area. By this means even the continental interiors are kept in isostatic equilibrium with the distant ocean basins. This implies a great depth and thickness to the zone of plastic flow. Although it must be plastic under moderate permanent stresses, this does not imply by any means a necessarily fluid condition, and fluidity is disproved by other lines of evidence.

The zone of compensation, being competent to sustain the stresses imposed by the topography and its isostatic compensation, must obey the laws pertaining to the elasticity of the solid state and is to be regarded therefore as of the nature of rock. Consequently there may be extended to all of it the name of the litho-

sphere, even though it includes from time to time molten bodies, the constituents of the pyrosphere.

The theory of isostasy shows that below the lithosphere there exists in contradistinction a thick earth-shell marked by a capacity to yield readily to long-enduring strains of limited magnitude. But if such a zone exists it must exercise a fundamental control in terrestrial mechanics, in deformations of both vertical and tangential nature. It is a real zone between the lithosphere above and the centrosphere below, both of which possess the strength to bear, without yielding, large and long-enduring strains. Its reality is not lessened because it blends on the limits into these neighboring spheres, nor because its limits will vary to some degree with the nature of the stresses brought upon it and to a large degree by the awakening and ascent of regional igneous activity. To give proper emphasis and avoid the repetition of descriptive clauses it needs a distinctive name. It may be the generating zone of the pyrosphere; it may be a sphere of unstable state, but this to a larger extent is hypothesis and the reason for choosing a name rests upon the definite part it seems to play in crustal dynamics. Its comparative weakness is in that connection its distinctive feature. It may then be called the sphere of weakness—*the asthenosphere*, and its position among the successive shells which make up the body of the earth is as follows:

The atmosphere The hydrosphere	Including the biosphere
The lithosphere The asthenosphere	Including the pyrosphere
The centrosphere, or barysphere	

Each has played its fundamental part in the development of earth-history.

STRESS-DIFFERENCES BETWEEN CONTIGUOUS COLUMNS OF THE CRUST

Stresses under conditions of isostatic equilibrium.—The continental platforms slope down into the ocean basins at grades which range mostly from one in ten to one in thirty. Some of the great

foredeeps show both the greatest depths of water and the steepest descents. The Chilean coast, for instance, at lat. 25° S., slopes from the Andes to a depth of 7,500 meters with a submarine grade of one in eight. Under the hypothesis of nearly perfect isostasy, which will be favored in this discussion, this would be taken to show the contiguity of areas in the crust of markedly unlike density.

Let the slope between such areas be regarded as a thick partition between two columns, each in isostatic equilibrium. These rest then upon the substratum below the zone of compensation with the same pressure and stand vertically in equilibrium.

In so far as the rock within the crust is subjected to mere cubic compression, equal in all directions and increasing with depth, there is no distortional force. In so far, however, as side pressures in one column are not balanced by equal side pressures from the adjacent columns, there is a stress-difference which does produce a distortional strain. If the stress-difference exceeds the elastic limit a permanent deformation results which reduces the stress and eases the strain. It is the plan of this paper to discuss the nature of the stress-differences on the partition separating two contiguous columns of the crust, of markedly unlike density; first, when these are in isostatic equilibrium, and second, when not in such equilibrium. Fig. 13 is drawn to show graphically these relations.

The land-column of the crust is marked *M;* the submarine column is *N; O* is the earth-shell below the zone of isostatic compensation; *P* is the column of sea-water. The vertical partition between the unlike columns stops in reality, according to the hypothesis, at the bottom of the columns. It is here extended down through the earth-shell *O–O* in order to discuss the deformation which would take place in the latter shell. *M* and *N* represent what is here called the lithosphere; *O–O* the zone which it is proposed to call the asthenosphere.

In case A, isostatic equilibrium is assumed and the pressures of the two lithospheric columns are equal upon the asthenosphere. But, assuming for the moment that the vertical pressures are freely transmitted as lateral pressures, it is seen that a marked horizontal unbalanced pressure is produced by the land-column against the

sea-column, as represented by the horizontal lines of the stress diagram. The top of the land-column is balanced only against the negligible weight of the atmosphere and the lateral stress gradient is there highest. The next portion below is balanced against the

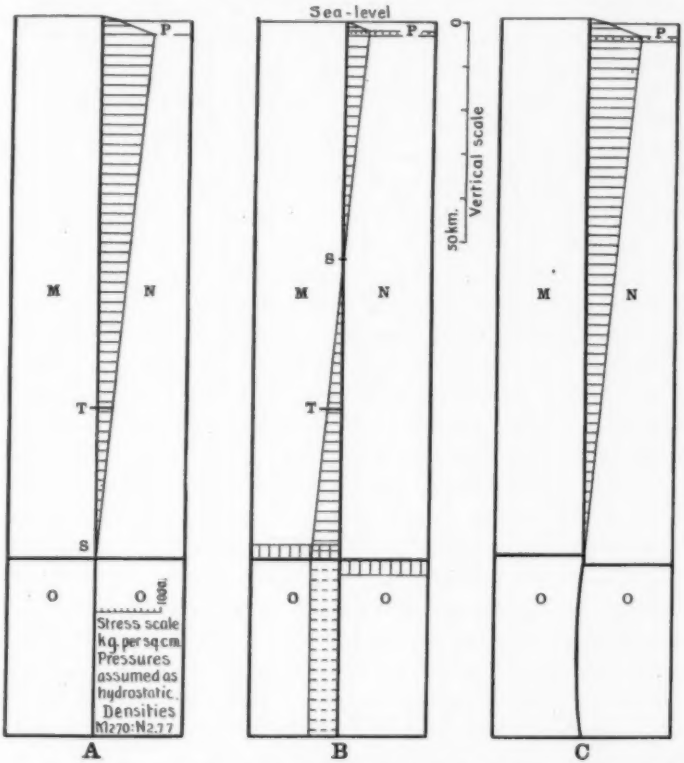

FIG. 13.—Diagram illustrating pressure-relations of the crust for marginal portions of the continental shelf and oceanic basin, interpreted as balanced by uniformly distributed isostatic compensation. Stress-differences are shown by cross-lined diagrams, the pressures being regarded as transmitted hydrostatically. The actual lateral stress-differences, for stresses within the elastic limit, are about one-fourth of the hydrostatic pressures here shown.

A. Columns in isostatic equilibrium.
B. Relations after base-leveling.
C. Relations after re-establishment of isostatic equilibrium.

sea-water and the stress gradient becomes less high. The maximum thrust occurs at the bottom of the ocean and is from the land toward the sea. Below this level the density of the sea-column is greater than that of the land-column. This, with increasing depth, gradually balances the excess pressure, and at the base of the lithosphere both the lateral and vertical pressures of both columns by hypothesis are equal.

In this diagram the pressures of the columns are imagined to act hydrostatically, but, in reality, for stresses *within the elastic limit*, this would not be so. Further, in so far as the partition is much wider than the difference in elevation of the columns, it has a gentle surface slope and will tend to give the upper part of the land-column competence to hold itself in by its own strength and that of the partition. The approximate ratio which the actual lateral pressure-differences on the two sides of the partition hold to the assumed hydrostatic pressures may be perceived from the results of a recent work by Love entitled *Some Problems of Geodynamics*.[1] In chaps. ii and iii he considers the problems of the isostatic support of continents and mountains. As a basis for the analytic treatment he assumes, first, the existence of complete compensation within a depth of one-fiftieth of the earth's radius, 127 km.; second, that at this depth all stress-differences disappear, the pressures below being of the nature of hydrostatic pressures, the only kind which could occur if a fluid layer existed at and below the depth of 127 km.; third, it is known that the heterogeneities of mass in the lithosphere only slightly modify the form of the geoid, and it is accordingly assumed that there is no such effect. Love thus treats of the limiting case of a crust exhibiting perfect isostasy, its surface relief not modifying the form of the geoid given by the ocean surface, and resting with its base upon a fluid zone. As such, his solution is of great value, but he states: "It must, however, be understood that the special form (of the hypothesis of isostasy) is introduced for the sake of analytical simplicity rather than physical appropriateness."[2]

The artificiality of the assumption of the existence of no stress-differences below the zone of compensation is shown by the law of

[1] Cambridge University Press, 1911. [2] *Op cit.*, p. 7.

density distribution which results. With only these three limiting assumptions, the number of unknown quantities remains larger than the number of equations, and the results are, strictly speaking, indeterminate; but by making various reasonable further assumptions definite solutions in accordance with these may be obtained. The elimination of stress-differences at the base of the lithosphere, taken as equivalent here to the zone of compensation, requires, however, that there shall be a peculiar relation of densities. To compensate an elevation it must be offset by matter below of less density than the mean for that depth, but in order to quench the stress-differences at the base of the lithosphere there must be between the light matter and this base a layer of more than mean density for that depth. Thus the light layer must perform a two-fold function, compensating not only the elevation above but the heavy layer below. For depressions in the crust there must be a reverse arrangement, matter of more than mean density existing immediately below the surface. But above the base of the lithosphere there must be a layer of less than mean density for that depth. The artificialities of this scheme would be sufficient to form a disproof of the initial assumption which determined it, but it also seems to be directly disproved by the evidence brought forward in the earlier parts of the present article. Nevertheless, the exact mathematical solution of this difficult problem is of great value as giving the results of the assumptions of extreme isostasy.

For the largest inequality of the crust, regarded as a zonal harmonic of the first order, that represented by the land and water hemispheres, Love shows that the lateral stress-differences under this hypothesis of isostasy reach a maximum at a depth equal to one-third of the zone of compensation and are equal to only 0.006 of the weight of a column of rock of height equal to the maximum height of the inequality. For harmonics of the second and third orders, representing the continents, the fractions are 0.0134 and 0.0208. These results, Love states, are extremely favorable to the hypothesis of isostasy, since the inequalities could be supported by any reasonably strong material.

There are two criticisms, however, to be noted while citing this conclusion. First, it is known that the crust is vastly stronger than

these requirements, so that such a perfected isostatic arrangement is not demanded on the score of crustal weakness. Second, the harmonic curves giving these figures are of a gently sweeping character; whereas, the actual continents are in many places high on their margins, and from these margins they slope with comparative steepness to the mean depth of the ocean floors. The stresses set up beneath the continental margins are accordingly a closer approximation to those imposed by lofty mountain ranges. Assume that compensations of the continental margins are perfect and the problem becomes that which Love takes up in the following chapter, namely, the isostatic support of mountains, except that we deal with only one great slope, whereas the theory calls for a succession of mountains and valleys.

It is shown that for such a compensated series, postulating the distribution of densities previously discussed, the greatest stress-difference exists at the mean surface, beneath the crests, and reaches a value equal to half the weight of a column of rock equal to half the height of the crests above the valley bottoms. From this maximum the stress-difference decreases to zero at the base of the zone of compensation. The solution by G. H. Darwin for uncompensated mountains and valleys gave a maximum stress-difference equal to 74 per cent of half the height, this maximum occurring at a depth equal to about one-sixth the distance between mountain crests. Even with perfect isostatic compensation, distributed after the fashion assumed by Love, the stress-differences for mountains and valleys are seen consequently to be two-thirds in value of those produced by an uncompensated relief, and are approximately one-fourth of the hydrostatic pressures. This fraction, one-fourth, happens also to be the same as Poisson's ratio, the ratio of the lateral expansion to the vertical shortening of a free rock column under vertical stress.

Now the distribution of density has been found to be more or less irregular, and there is no evidence of such a reversing layer at the base as Love has postulated. Stress-differences will consequently extend below the isostatic compensation. If, however, the latter is not uniformly distributed, but is concentrated somewhat in the outer half of the lithosphere, the stress-differences will become

small at and below the base of the lithosphere. On account of the incompleteness of local compensation, the irregularities and uncertainties of the actual facts of nature, the Gordian knot of a solution may be cut by simply assuming for present purposes the form of diagram given by hydrostatic pressures due to a compensation uniformly distributed. The approximate stress-differences will be given by taking one-fourth of the values given by the hydrostatic pressures. This transfers the problem from the difficult field of zonal harmonics to the simple one of hydrostatics, and perhaps does not introduce errors greater than those involved in the differences between nature and the postulates which form the foundation of the solution by zonal harmonics. This hydrostatic diagram is shown accordingly in Fig. 13.

It is held by the advocates of extreme isostasy, however, that for long-continued stresses the crust is very weak; in other words, the elastic limit is low, and slow plastic deformation readily occurs which tends to dissipate the stress-differences and re-establish isostatic equilibrium. To the extent to which this is true, the real diagram of lateral stresses would approach the hydrostatic diagram here given and measure the forces producing plastic flow.

It has remained, however, for the opponents of the hypothesis of local and nearly perfect isostasy to point out, what is here illustrated graphically, that the extreme theory requires a belief in vertical weakness but lateral strength. If it were not for lateral strength the land-column would crowd against the sea-column, more at the top than at the bottom. Flowing out with a glacier-like motion over the upper part of the sea-column, the land-column would settle at the top and become shorter. This in turn would bring about a vertical elevatory pressure against its bottom, the column would rise, lateral creep would continue with equal pace, and the end result would be a density stratification in which the continental crust would come to overlie the oceanic crust. The limit of such an action would be given by the decreasing surface gradient, this finally becoming so gentle as to stop the glacier-like flow. The lack of such an effect implies of course that the lateral stresses of the outer part of the lithosphere lie within the elastic

limit. Therefore they may be regarded as having not more than a quarter of the value shown in Fig. 13A.

The suggestion of the existence of opposing modifying factors is to be found in conclusions from the previous parts of this investigation—that compensation may be in many places concentrated somewhat in the outer half of the zone as here shown and in other places fade out through a notable distance below. These two variations in the distribution of compensation would modify the stress diagram in opposite directions.

Modifications of stresses produced by base-leveling.—Consider next the case of complete erosion to sea-level, as shown in Fig. 13B. The rock from the land-column has been deposited as sediment over the sea-column. As the columns are supposed to act as units the sediment is shown as spread uniformly. The lateral stress diagram beneath the bottom of the sediment shows a rate of decrease the same as in case A, but the value of the hydrostatic stress at any depth is diminished by the sum of the depths of erosion and deposition. The lateral stress now changes in sign at a point S and at this depth is a line of no lateral stress. Above this depth the continental segment tends still to spread over the ocean, but less effectively than before; below this depth the oceanic segment now thrusts against the continental crust.

If the ocean water be eliminated from the diagram and base-leveling should bring both columns to a uniform surface, then the neutral depth S advances to the surface and the lateral stress diagram in B is just the reverse in value to A. In that case there is no lateral thrust at the surface, but at all depths below there is an excess pressure against the continent reaching a maximum at the bottom of the lithosphere. This extreme case cannot apply to the ocean except for that limited width over which is built out a continental shelf. To the degree to which the weight of this shelf is supported by the ocean crust beyond, the column beneath the shelf would not operate with its full pressures against the land. The case would apply better to the complete erosion of level-topped plateaus situated within a continent.

For the lateral pressure within the lithosphere to become effective in a landward undertow would require a lesser rigidity of the

crust at the bottom than at the top. Such a lesser rigidity may be granted, but it is seen then that the landward undertow would be greatest at the bottom and could not advance above a depth indicated on the diagrams by T. At this point the stress is of the opposite sign but of the same value as for the state of isostatic balance in case A. If seaward flow did not take place at this level in the first case, landward flow could not take place in the second.

For the extreme case where isostasy is completely destroyed by surface leveling, no water body remaining, T will rise upward to a depth equal to one-half the depth of the zone of compensation. If the surface of complete compensation be 76 miles deep, this gives a minimum depth of 36 miles. For the undertow to reach this height implies, however, not only the limiting case of complete destruction of isostasy, but a crust only one-half as rigid at depth T as at the surface and a previous state of expansive surface stress as great as the outer crust could bear. On the other hand, if tangential pressures due to centrospheric shrinkage should co-operate with the stresses tending to restore isostatic equilibrium, underthrust would become more effective below, but overthrust would also become effective above.

The disappearance of isostatic compensation at a certain level means the disappearance of notable heterogeneity in the earth-shell below, as argued in Part V (pp. 446–48). One of the possible suppositions to explain this is to suppose that this shell is weaker than the crust above and therefore the lateral thrust due to an assumed initial heterogeneity would cause a lateral flow, a density stratification, and a resulting disappearance of the postulated heterogeneity. This supposition of a weaker zone finds support in other lines of evidence. Therefore, although some lateral flow at the base of the lithosphere may occur during the restoration of isostatic equilibrium, it is to be expected that the bulk of such flow will be below, for there the substance is more plastic and the lateral stress is throughout at a maximum.

Let attention be given next to the vertical as contrasted to the lateral unbalancing brought in by the destruction of isostatic equilibrium. The land-column becomes lighter, the sea-column heavier, by amounts which are shown in the *vertically lined* stress

diagrams at the base of the lithosphere in case B. Supposing that vertical readjustment of the columns is prevented for a time by the strength of the crust, the vertical stresses will be taken up by a vertical shearing strain along the partition between the two columns. This shear is equal in amount to the difference in total weights of the columns. Let the shear per unit area be called s. It acts over a surface taken as 122 km. high. Let this height be called h. The weight of the columns will vary with their breadth in the plane of the drawing. If the breadth of each be taken as b and the weights per unit area as M and N (N including rock, sediment, and sea-water), then for a cross-section of unit thickness the total difference in weight is $(N-M)b$ and the total shear is this same amount, provided that the columns are not sustained in part by other boundaries. But the total shear is also sh. Therefore

$$sh=(N-M)b$$

$$s=(N-M)\frac{b}{h}$$

For narrow columns b is small, giving to s a small value and consequently one within the elastic limit. Let b become broad and s will then become large and exceed the elastic limit. The *lateral* pressures, on the contrary, are less dependent upon the breadth, and, if the problem were regarded as one of hydrostatic pressures, would be wholly independent of breadth. The formula shows that the broader the columns, the more readily they will readjust by vertical shear between the columns. Now unless failure by vertical shear took place between the upper part of the columns the heavy column would be held up, the light column would be held down, except for the partial effect of sagging in case the columns were very broad. The lateral landward pressure at the base could therefore not become effective. The loaded portion of the crust must fail first by shear or flexure of its upper portion. Whatever be the distribution of strength it would appear then that the primary yielding is the vertical one and the landward force of undertow can become only secondarily effective.

The hypothesis of local and nearly complete isostasy requires that the elastic limit for vertical shear should be very low in order

that narrow columns should be able to rise or sink. This may be illustrated by the following example:

Suppose a region 50 km. in radius possesses a mean departure from isostatic equilibrium equal to 76 m. of rock (250 ft.) and that the surrounding regions are out of adjustment by the same amount but in the reverse direction. This is the maximum area for regional isostasy which in Hayford's opinion is to be expected, and 250 ft. is the mean departure from isostasy as given by him in his Minneapolis address. But in this example the adjacent regions are each assumed to be out of adjustment in opposite directions by this amount and, therefore, the differential load is twice this or 500 ft. of rock. The case is one which he would regard consequently as rather extreme. Now a cylinder 100 km. in diameter and 122 km. deep could not fail through its bending moment, as in the flexing of a beam. It would have to fail as in punching a rivet hole through a metal plate, in other words, by circumferential shear. The shearing stress per unit area is obtained by dividing the total load by the total shearing surface. With the data taken as above this gives $s = 8.4$ kg. per sq. cm. or 120 lbs. per sq. in. But strong rock at the surface can readily carry a shearing stress of from 700 to 1,000 kg. per sq. cm. (10,000 to 14,000 lbs. per sq. in.). Isostatic perfection to this degree would therefore require the zone of compensation as a whole to be only about one-hundredth as strong under permanent stress as is solid rock at the surface. This calculation alone would tend to show that the loads and areas by which the crust departs from isostatic equilibrium have been much underestimated by the advocates of extreme isostasy.

It should be noted, however, that, following the lines of his rejoinder to Lewis, Hayford would answer that he regarded the landward isostatic flow as taking place within the zone of isostatic compensation and the vertical shear as operating, consequently, through a depth far less than the thickness of the entire zone of compensation. There are, however, a number of inconsistencies in this argument, some of which have already been made evident. Others will appear as a result of the later discussion of this chapter. But it may be noted that even granting this contention—that only the outer third of the zone of compensation was involved—the

unit shearing stress would be multiplied only by two or three and would still imply a weakness in this part of the crust to resist long-enduring shear or bending stresses, its capacity being only 3 or 5 per cent at most as great as is found to exist in surface rocks for stresses of human duration.

Relief of stress accompanying restoration of isostasy.—It is seen from the preceding analysis that the movement of the unbalanced columns toward a new state of equilibrium will be partly by vertical shear in the neutral ground between them, but, where the areas are large in comparison with the thickness of the zone of compensation, the easiest mode of yielding may be by flexure, showing at the surface as crustal warping. Both modes of yielding serve to transmit the excess vertical stresses of the heavy and sinking column into the asthenosphere. If the latter be indeed a shell of weakness it will transmit these pressures more or less hydrostatically. The vertical pressure-differences will act within it as lateral pressures making for flow toward the lighter column. It is shown in Fig. 13B that the maximum horizontal stress in so far as it approaches a hydrostatic distribution acts throughout the whole depth of this zone, so that it not only is weaker than the crust above, but is subjected to maximum stress over a greater area. It will yield by flowage therefore either if of small depth and very plastic, or of great depth but more rigid. If the columns are adjacent and narrow as compared to the thickness of the shell of weakness, then the principles of plastic flow would require that the flow be chiefly in the upper part of this shell. If, however, the columns are of considerable breadth compared to the thickness of the asthenosphere, and especially if at a distance from each other, then the principle of least work would determine that the middle strata of this shell should flow the farthest and the whole would to some degree participate. If an imaginary partition were extended downward through this shell as shown in A and B of Fig. 13 this partition would be found warped after the movement as shown in C of the same figure.

It was seen in an earlier part of this discussion that, even supposing deformation became effective by means of the lateral stresses within the lithosphere and without the existence of a zone

of weakness below, still only the basal part below the point *T* would be competent to give a landward movement during the restoration of isostatic equilibrium. But now it is seen that in the asthenosphere the lateral pressures are transmitted with greater amount, from a greater distance, and with a greater cross-section. The zone is one without notable isostatic compensation within it and is presumably more plastic than the basal part of the lithosphere. Therefore there is good reason to believe that the subcrustal undertow is restricted to the asthenosphere.

The forces actually needed to produce flowage would be in reality but a fraction of those indicated in Fig. 13B as existing in the asthenosphere. The reason is that the greater part of the vertical forces is consumed in producing flexure and shear in the lithosphere. Only a residuum is needed to produce a slow plastic flow in the shell below. For that reason broken lines are used in that part of the stress diagram. The energy consumed within the lithosphere by its deformation will be nearly independent of the breadth of the columns; it will actually tend to become somewhat less with breadth because flexure on large radii will be favored. The energy consumed in the asthenosphere will, on the other hand, increase with the breadth of the columns, but will be spread over a greater area. The temperature effect due to the absorption of energy would appear to be a minor factor, for it cannot exceed that energy which is supplied by the average vertical stress-difference multiplied by the vertical distance moved. The average vertical stress-difference will be the mean between that at the beginning of movement and that residual stress remaining after the movement is completed.

In determining the scale of the diagrams of Fig. 13 the following data were chosen. The land-column was taken in A as having a surface elevation of 1,000 m. and a density of 2.70; the sea as 3,000 m. deep, and the rock below as possessing a density of 2.77. The sea-water has a density of 1.03. These relations give an isostatic balance at a depth of 122 km. In B, erosion of the land to sea-level is supposed to have taken place and the sediment spread with same unit weight over the sea-column that it had as

rock upon the land. These relations give a depth of 54 km. to *S* and 88 km. to *T*.

It should be repeated, however, in closing this topic, that the solutions here given are approximate only and assume that isostatic compensation results in lateral stress-differences which show the same distribution of forces as a diagram of hydrostatic pressures, differing only in magnitude. The writer is inclined to think that the actual facts of nature call in most cases for some depression in depth of the critical points beyond those here shown. Especially is there likely to be under the margins of a continent in isostatic equilibrium some permanent lateral stress-difference within the asthenosphere, due to the compensation above and tending toward a landward undertow. Upon the unbalancing due to erosion and sedimentation this would cause the lateral stress-differences within the asthenosphere to rise more quickly to the low elastic limit and permit more readily than would otherwise be the case a regional readjustment toward isostasy.

RELATIONS OF UNDERTOW TO THE ZONE OF COMPENSATION

Present status of the problem.—The causes of vertical movements Dutton[1] made twofold. He clearly distinguished on the one hand between those internal forces leading to expansion or contraction, which tend, by producing changes in density, to create isostatically a new surface relief, and, on the other hand, those isostatic readjustments following erosion and sedimentation, readjustments which tend not to make a new, but to restore the older, relief. Folding he regarded as unrelated to the former, as a result of the latter. He had shown earlier (in fact, he had the honor of being the first to show) that the time-sanctioned hypothesis of cooling as a cause of crustal shrinkage and consequent mountain-making was inadequate to account for either the distribution or amount of folding.[2] From this he was led to regard folding as due, not to any kind of contraction, but as a compressive movement of one section

[1] "On Some of the Greater Problems of Physical Geology," *Bull. Phil. Soc. Wash.*, XI (1889), 51–64.

[2] C. E. Dutton, "A Criticism upon the Contractional Hypothesis," *Am. Jour. Sci.*, VIII (1874), 113–23.

of the crust against another, presumably offset by tension in some other region. Dutton's argument is that the crust beneath the plateau is unloaded by erosion, that the crust beneath the basin is loaded by sedimentation. An isostatic movement, rejuvenating the relief, must, by causing the overloaded basin to settle, produce a squeezing-out of matter beneath the sinking area, and a crowding-in of matter beneath the rising area. The surficial movement of sediment is from the high area toward the low. The deep-seated movement is from the low toward the high. Thus the cycle becomes completed and the mass of matter above the level of complete compensation remains the same in each column. The seaward movement of the sediment, as a frictional resistance against the river bottoms, produces only an insignificant drag, but the return subterranean movements by viscous or solid flowage must produce a pronounced drag upon the crust in the direction of the rising region. Dutton's reasoning is clear, but the effectiveness of the action rests upon several assumptions. First, it omits the influence of the surface relief and the degree to which that tends to a lateral spreading movement from the high toward the low regions. Secondly, it postulates a low rigidity to the crust, as he in fact notes. Thirdly, it involves the conception of a strong undertow fairly near the surface in order that the crust above may be too weak to resist the viscous drag. As there were little quantitative data available at the time when Dutton formulated this corollary of his theory of isostasy he could not have tested the validity of these assumptions, but raised the problem for those who should come after him.

This theory of folding took a somewhat different form in the mind of Willis, as expressed in the concluding chapter of his *Research in China*.[1] This work in many ways is of the very first importance and gives a comprehensive view of the geological history of the whole continent of Asia. As to the nature of the movements, he finds that the continent of Asia may be resolved into positive and negative elements, the former areas tending to stand high, the latter tending to stand low. These tendencies are latent during comparatively long periods of quiet and resultant peneplanation,

[1] Vol. II (1907), Carnegie Institution of Washington.

but become operative during epochs of diastrophism. The compressive movements, on the other hand, have pressed and welded the positive elements together, the axial directions of folding representing the compression of the negative zones lying between.

The cause of the diastrophism Willis ascribes to differences in specific gravity, restricted, according to Hayford's determination, to the outer hundred miles of the earth's body; the vertical movements being chiefly due to isostatic readjustment between the several continental elements, the compressive movements being due to the tendency of the heavier oceanic segments of the earth to spread and underthrust the outer portions of the whole continental mass. This theory of the cause of lateral compression was discussed by the present writer in a review of Willis' work,[1] and the objections stated against it there are in part the same as will be elaborated farther on in the present article.

Hayford took up the same subject in his address, delivered at Minneapolis on December 29, 1910, as retiring vice-president of Section D (Mechanical Science and Engineering) of the American Association for the Advancement of Science, the title of his paper being "The Relations of Isostasy to Geodesy, Geophysics, and Geology."[2] This is a paper of broad scope intended to show how vertical movements not in apparent accord with isostasy and also movements of folding may be explained as secondary results of isostatic adjustment and really in harmony with the hypothesis of nearly continuous movement in a crust of low rigidity and of almost complete isostasy. This part of his theory is essentially the same as Dutton's but is elaborated in greater detail.

Harmon Lewis called attention to the defects in this theory of deformation,[3] but Hayford made a rejoinder, positive and sweeping in its style, to this and other lines of criticism by Lewis.[4]

The names of Dutton, Willis, and Hayford deservedly carry much weight and must be accepted at their face value by geologists

[1] *Science*, N.S., XXIX (1909), 257–60.

[2] Published in *Science*, N.S., XXXIII (1911), 199–208.

[3] "The Theory of Isostasy," *Jour. Geol.*, XIX (1911), 620–23.

[4] John F. Hayford, "Isostasy, a Rejoinder to the Article by Harmon Lewis," *Jour. Geol.*, XX (1912), 562–78.

who have not themselves made a critical study of the problems of isostasy. The arguments which the writer advanced in 1909 against this hypothesis were published under a title which apparently did not call attention to them. The style of Hayford's reply to Lewis is crushing and conveys the impression that Lewis has been completely refuted. It is because of these reasons that the subject calls here for fuller development.

In his Minneapolis address Hayford outlines a theory of the principles of diastrophism which turns upon his conclusion that isostasy is so nearly complete that areas of even limited size average only 250 feet from the level of isostatic equilibrium. He assumes chemical and physical changes to be induced in the crust by the changing load due to erosion and sedimentation. These he thinks are superimposed upon the effects of nearly continuous vertical movements of isostatic readjustment. The vertical movements in turn produce a lateral undertow which is given as a cause of localized heating and folding. Apparently this is regarded as a complete mechanism of deformation since the author raises the query:

> Is it at all certain that under the influence of such actions the geological record at the earth's surface at the end of fifty to one hundred million years would be appreciably less complicated than the geologic record which is actually before us? I think that it would be fully as complicated as the actual record.[1]

This theory of folding as the result of subcrustal undertow is illustrated by means of two diagrams. In Fig. 1, the zone of viscous flow from the sinking toward the rising area is placed in the lower quarter of the zone of isostatic compensation. In Fig. 2 it is shown in the middle of that zone, dying out both above and below. Apparently then, as shown by these two different conceptions, the author cited was guided by no definite theory, based upon the mechanics of materials, as to the factors which would determine the depth of this zone of undertow and its relations to the zone of compensation.

Harmon Lewis in his paper on the "Theory of Isostasy" has discussed various aspects of the isostatic theory as developed by

[1] *Op. cit.*, p. 206.

Hayford, and among them this question. Regarding the possibility of folding by means of isostatic undertow, Lewis concludes:

> Now, according to the theory of isostasy, compensation would be essentially complete, and if compensation is complete the depth of compensation as determined by Hayford's geodetic work would be as great as 60 miles. Hence, the undertow postulated by isostasy would exist chiefly below 60 miles. It is decidedly questionable that an undertow even much nearer to the surface than 60 miles would cause the observed folding in the upper few miles of the crust.[1]

In regard to this criticism by Lewis concerning the cause of folding, Hayford states in reply:

> On pp. 621–22 Mr. Lewis sets forth the argument that there is much geological evidence of horizontal movements in the outside portions of the earth, especially in the form of folding, that the controlling movements of isostasy are assumed to be vertical and hence cannot account for folding, and that the horizontal movement or undertow concerned in isostatic readjustment must be below the depth of compensation and hence so far below the surface as to be very ineffective in producing folding.
>
> There are two fatal defects in this argument as applied to controverting anything that Hayford believes or has written.
>
> First, Hayford has already indicated clearly his belief that the undertow concerned in isostatic readjustment is above, not below, the depth of compensation. In both the figures published in his Minneapolis address the undertow is clearly indicated as being above the depth of compensation and it is also so indicated in the corresponding text. As Hayford puts the undertow comparatively near the surface, where it is conceded that it would be effective in producing folding, the existence of extensive folding is a confirmation, not a contradiction, of his theory of the manner in which isostatic readjustment takes place. It is certainly not fair to hold Hayford responsible, either directly or by inference, for any theory which someone else may believe which involves an undertow situated entirely below the depth of compensation. Mr. Lewis apparently believes such a theory.
>
> Second, the movements which produce isostatic readjustment are necessarily horizontal, not vertical. If two adjacent columns of the same horizontal cross-section extending from the surface to the depth of compensation have different masses the readjustment to perfect compensation must involve a transfer of mass out of one column, or into the other, or from one to the other. In any case the transfer must be a horizontal movement. Hayford has already shown in print more than once that he understands that vertical movement alone does not produce isostatic readjustment. Moreover, a careful reading

[1] *Op. cit.*, p. 622.

of his Minneapolis address will certainly show that he believes that the total amount of material moved horizontally during isostatic readjustment, and especially the total number of ton-miles of such movement, is vastly in excess of the corresponding quantities concerned in the vertical components of the movement which takes place. Hence the folding and other abundant evidence of past horizontal movements observed by geologists confirm Hayford's hypothesis as to the manner in which isostatic readjustment takes place, instead of conflicting with it as Mr. Lewis' article would lead one to think.[1]

The present writer, however, believes with Mr. Lewis in the theory that an undertow must be essentially below the zone of compensation and is incapable of producing surficial folding. The reasons have been given in part in the consideration of the stress-relations, as they would exist under the hypothesis of extreme isostasy. But there are other reasons why the subject should be discussed in further detail. One reason is that, if Lewis is right on this point and Hayford wrong, it is desirable that this should be made clear, in justice to Mr. Lewis as well as to the subject. The other reason is that here in reaching a conclusion we can advantageously pursue a method of exclusion. By showing that isostatic undertow cannot take place within the zone of compensation, for various reasons besides those discussed in the stress diagrams, we reach the conclusion that it must take place in a level below that zone. Furthermore, by noting the conditions which would hinder lateral flowage we may arrive at a conclusion as to those which must exist to greater or less degree in order to permit it.

Objections against undertow in the zone of compensation.—The pressures which occur during a state of isostasy and after the destruction of that condition have been discussed. It was seen that the pressures making for the undertow necessary to restore isostasy were greatest at the bottom, but, more especially, below the bottom of the zone of compensation. The possibility remains to be considered, however, that perhaps the distribution of the rigidity of the crust more than offsets the distribution of pressures. Suppose the middle of the zone of compensation should be very weak and the crust at and below the bottom be very strong. Then,

[1] *Op. cit.*, pp. 573, 574.

if the restoration of isostasy was deferred until assisted by strong tangential pressures due to centrospheric shrinkage, it might be held that isostatic undertow could take place within the zone of compensation and between *S* and *T* of Fig. 13B. If, furthermore, compensation should be not uniformly distributed but taken as largely concentrated in the upper part of the zone of compensation, which however is contrary to the Hayfordian hypothesis, then the forces making for undertow may correspondingly rise in the crust. For these reasons it is seen that the previous argument from the distribution of pressures is not final and that the physical conditions involved in lateral flowage must also be considered.

The only positive reason which has been advanced for seeking to place the undertow within the zone of compensation is in order to utilize its viscous drag as a cause of folding. To become effective the drag must be strong, the crust above by contrast weak and therefore thin. The crumpling pressure on the *surface* of the crust cannot be transmitted directly from the sinking area, as is shown in Fig. 13, since the thrusting force is greatest at the bottom. It must be supposed to arise from the viscous drag of the undertow. But viscosity decreases the hydrostatic head with increasing distance from the source. Therefore, to permit a viscous flow at a distance from the source of pressure implies a mobility within that level of the crust which would make it wholly incapable of carrying the stresses necessary to maintain its own isostatic equilibrium. Therefore this level, by the very terms of the general conception of isostasy, would become the bottom of the zone of compensation.

As another mechanical defect of the theory under review, it is to be noted that the section of undertow taken by Hayford as in the middle of the zone of compensation is not given as involving more than half of that zone. This is as if a viscous fluid were transmitted through a pipe in which the cross-section of pipe and fluid were equal. To assume that the fluid is free to escape into a region of less pressure at the far end and yet gives such a frictional resistance against its walls as to be able to crumple up the pipe is to assume that the two are of the same order of strength. The materials of pipe and fluid might almost be interchanged.

In such viscous flow the tendency would be for a swelling and bursting to appear at the near end rather than a through flowage with a crumpling of the pipe at the far end.

Finally, the greatest theoretical difficulty is encountered when it is sought to transmit matter from beneath the regions of oceanic marginal sedimentation to beneath the regions of a continental interior. Either directly or indirectly there must be a subcrustal transference going forward all the way between these distant regions; for example, from beneath the Mississippi and Colorado deltas to the fields of erosion in the Rocky Mountains, if a condition of even approximate isostasy is to be maintained throughout. This does not mean of course that an individual ton of plastic rock is transferred a thousand miles to balance a ton of sediment. Each subcrustal unit may be transferred only a mile, but it involves a subsurface movement of matter all the way from the regions of sedimentation to the regions of erosion.

Now this implies a *continuous pressure-gradient*, and even under the conception of great crustal weakness, a pressure-gradient which could fold the weak cover-rocks would be far higher than that needed for the movement of a continental glacier. Any large degree of viscous resistance in the zone of undertow would therefore require, in order to initiate movement, an enormous defect of isostasy under the distant continental interior, an enormous excess under the marginal oceans. After a rejuvenative movement had started, it would be slow, the frictional and deformative resistances nearly balancing the deforming force. Therefore inertia of the moving mass could not carry it appreciably beyond the point where the moving force, weakened by loss of head, would just balance the resistances to further movement. It would be expected, in consequence, that a residual pressure-difference would remain, even after a period of restorative isostatic movement. But an inspection of the map of New Method anomalies given in Part II, p. 153, does not show any such anomaly gradients as would comport with this expectation. A vast region of the continental interior extending from Lake Superior to the Rio Grande and westward to beyond the front ranges of the Rocky Mountains shows average positive anomalies, indicating an excess of matter, not a

deficiency. To the westward is a broad region of average negative anomaly reaching a maximum at centers near the Pacific coast and no marked excess is shown near the mouths of the great rivers. Such a lack of regional relations would appear to show that the anomalies are due much more to local loads and irregularities upon and within the lithosphere, and to bowings due to great compressive movements unrelated to isostasy, rather than to the existence of an isostatic gradient leading from the ocean borders to the interior fields of great erosion. Therefore either the idea of strong viscous drag by undertow or else the very doctrine of isostasy—one or the other—must be abandoned. But it has been seen that if undertow exists in a comparatively plastic stratum, then that physical condition will cause it to be the bottom of the zone of compensation. Thus the application of every pertinent engineering principle reduces the initial hypothesis of surface folding by isostatic undertow, and, especially by undertow within the zone of compensation, to an absurdity.

Undertow restricted to a sphere of weakness—the asthenosphere.—All of this accumulative argument has not been advanced merely to show that a certain view is wrong. Rather has it been the intention to prepare the ground for what would appear to be a sounder theory of the mode of maintenance of isostatic equilibrium.

As for the basis of that theory, Schweydar, from the mathematical analysis of the measurement of the tides in the crust by means of the horizontal pendulum, has found that they are in accord with the assumption of the existence of a slightly plastic zone about 600 km. thick beneath a more rigid crust 120 km. thick.[1] It would appear that the geodetic evidence of isostasy points also toward the existence of such a thick and somewhat plastic zone beneath the more rigid lithosphere. It gives no knowledge of the exact thickness or depth, but for convenience the figures given by Schweydar will be assumed. It is a matter of importance to note however that, although the quantitative limits are uncertain, the suggestions given both by the tides and by isostatic

[1] "Untersuchungen über die Gezeiten der festen Erde und die hypothetische Magmaschicht," *Veröffentlichung des k. k. Preusz. geodät. Institutes*, Neue Folge No. 54, Leipzig (1912, B. G. Teubner).

compensation point to a zone of weakness much deeper and thicker than the figures which have customarily been taken as a probable depth of origin of magmas. The latter however rests upon uncertain extrapolation, whereas the figures for the limits of the asthenosphere, although of no exactness and perhaps 20 or 50 per cent from limits which finally may be chosen, have at least been determined by more direct evidence. In such a thick shell of weakness, the readjustment, after an erosion cycle, of a continental interior to isostatic equilibrium would require but very little viscous shear and but little lateral movement.

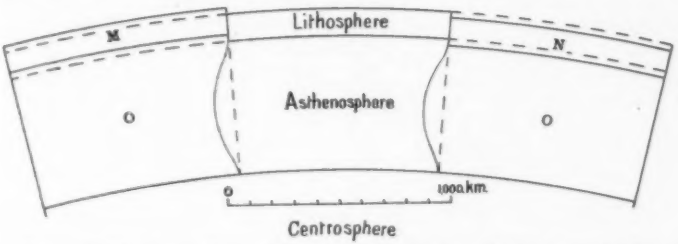

FIG. 14.—Diagrammatic vertical section of the crust, to show nature of undertow in the asthenosphere necessary to restore isostatic equilibrium in a positive interior continental area after a cycle of erosion. Effects of a vertical movement of 0.5 km. exaggerated 60 times. Asthenosphere grades into contiguous spheres and best limitations in depth are not known.

To give quantitative visualization to this conclusion Fig. 14 is drawn. Suppose a plateau area 1,000 km. wide in a continental interior to be separated from the region of sedimentary deposit by an intermediate region 1,000 km. across. Take a section 1 km. wide through these regions. Let an erosion cycle cause the removal on the average of 0.5 km. of rock from this area to be deposited over an equal area of sea-bottom. Then, during an epoch of diastrophism, assume complete recovery of isostatic equilibrium by undertow in a sublithospheric zone of weakness 600 km. thick. The vertical section of rock eroded is 500 sq. km. in area. As we have chosen a width of section of 1 km. we may also speak of this as the volume, 500 cu. km. To restore the mass of this column, 500 cu. km. must be added to it and flow past the vertical line which bounds it on the seaward side. As this zone of

flow is 600 km. deep, the actual lateral movement, if all depths move equally, will be but 0.83 km., since $0.83 \times 600 = 500$. If the flowage is supposed to increase regularly from top and bottom to the middle the movement of the middle layer would be 1.66 km. A previously vertical line 600 miles long through this asthenosphere would then be bent at the middle by this amount and its two halves make angles of $0°19'$ with the vertical. Each layer a kilometer thick would move horizontally 5.6 m. with respect to each adjacent layer of kilometer thickness. These figures bring out the insignificant degree of the plastic deformation in such a deep zone which is needed to restore isostatic equilibrium, even for a large interior continental area after erosion amounting to two-thirds of the present average elevation of the North American continent.

As a matter of fact the cross-section of the plastic deformation would not be a triangle, but a sinusoidal curve, so that the maximum linear flow for thickness of 600 km. would be between 0.83 and 1.66 km.

This illustration makes it clear that the isostatic rejuvenation of continental interiors as well as of the margins, which meets such grave difficulties under the hypothesis of a thin and shallow zone of isostatic undertow, is eliminated by adopting the hypothesis of a thick and plastic sublithospheric shell, such as has been found to be suggested by independent evidence.

The idea of folding as a result of isostatic undertow definitely may be abandoned, but the absence of a notable isostatic gradient has some further significance. It is seen from Fig. 14 that if the fields of great erosion and deposition are within a few hundred kilometers of each other the rejuvenative undertow, under the laws of stress distribution in plastic bodies, would involve mostly a limited tract in the outer part of the asthenosphere; whereas, if the undertow must extend over distances of 1,000 km. or more, then the whole depth of the asthenosphere will become involved. The amount of stress-difference and of plastic shear per unit of volume may therefore be no greater in the one case than in the other. Especially, if the middle of the asthenosphere is its weakest part, a movement generated by areas large enough to involve the whole of this zone would go forward under less stress-difference per unit

of area than for more local adjustments. The absence of a notable continental gradient is suggestive therefore of a deep zone of weakness, least resisting in its central portions, and of very marked plasticity in comparison with the rigidity of the lithosphere above. This does not involve, however, the conception of a truly fluid zone, but merely that of a comparatively plastic solid.

The existence and nature of this zone of weakness is seen to enter vitally into the theory of isostasy and must of course bear with equal importance on other branches of terrestrial dynamics as well. It is proposed therefore to elevate it to equal rank with the other shells of the earth and to name it for that quality which, from the standpoint of diastrophism, is its most significant feature as compared to the zones above and below. This is its inability to resist stress-differences above a certain small limit. Its name, therefore, is the sphere of weakness—the asthenosphere.

[*To be continued*]

A COMPARISON OF THE CŒUR D'ALENE MONZONITE WITH OTHER PLUTONIC ROCKS OF IDAHO

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The rough mountainous area comprising nearly all of the central part of Idaho and adjacent portions of Montana is an enormous batholith of quartz-monzonite of late Mesozoic age. In the northern part or "panhandle" of the state the predominating Algonkian sediments are cut by irregular masses of acidic intrusives which are quite generally believed to be outliers of the great batholith to the south. It is the purpose of this paper to point out that among these intrusives of northern Idaho is a type distinctly different from the other acidic intrusives of the state, and to propound for future field-workers the problem of its relationship to them.

The quartz-monzonites of central Idaho and those found in the southern part of the state are light-gray, granular rocks with biotite the chief dark silicate, and often containing muscovite. Porphyritic development of the feldspar and variations in the relative amounts of the minerals are common, but taken as a whole they are remarkably uniform rocks. I have seen this quartz-monzonite in place in the vicinity of Moscow, of Elk City, and in the Dixie mining district, and have examined specimens of it from other localities, and in every case it showed the same general characters. Professor D. C. Livingston has traversed great areas of it along the Salmon River; and the various United States Geological Survey reports by Lindgren, Eldridge, and Umpleby all describe the same light-gray, *micaceous* quartz-monzonite of which some of us have come to speak familiarly as the "Idaho granite." In the extreme north of the state around Priest Lake I have crossed many miles of the same rock.

Indeed it would be difficult in a pile of mixed specimens from these localities to distinguish the rocks from the various districts. Should there be in this pile, however, a sample of the average

intrusive from the Cœur d'Alene district, the most casual observer would note a difference. This rock, according to Calkins,[1] is

> a medium-grained granular rock of which the constituents megascopically recognizable are feldspar, hornblende, and a little quartz. Of the feldspars, striated plagioclase, opaque and white, in rather small idiomorphic crystals can be distinguished from somewhat larger, grayish, and more transparent imperfectly formed crystals of alkali feldspar. The two are present in nearly equal amounts. The microscope shows, in addition to the constituents named, a little biotite, some small prisms of light-green monoclinic pyroxene with titanite and magnetite as abundant accessories.

Variations from this type are noted, especially in respect to the relative amounts of the minerals, but none of them resembles the rock from central Idaho, from which all of the Cœur d'Alene monzonite differs in the smaller amount of quartz, in the presence of hornblende, and in the lack of megascopic mica. The chemical differences are indicated by Table I.

TABLE I

ANALYSES OF INTRUSIVE ROCKS FROM IDAHO

	A	B	C	D	E
SiO_2	61.41	72.07	68.42	69.56	65.23
Al_2O_3	17.99	15.51	15.01	15.29	16.94
Fe_2O_3	2.93	0.31	0.97	0.86	1.60
FeO	1.39	1.01	1.93	2.06	1.91
MgO	1.30	0.35	1.21	0.69	1.31
CaO	4.75	1.93	2.60	2.81	3.85
Na_2O	4.01	4.02	3.23	3.97	3.57
K_2O	4.59	4.09	4.25	3.36	3.02
H_2O-	0.11	0.03	0.54	.....	0.18
H_2O+	0.68	0.30	0.73	0.86	0.88
TiO_2	0.53	0.16	0.50	0.55	0.66
P_2O_5	0.19	0.11	0.13	0.16	0.19
All other	0.46	.....	0.38	.....	0.44
Total	100.24	99.89	99.95	100.17	99.78

A. Quartz-monzonite, near Gem, Idaho (Cœur d'Alene).—*Prof. Paper 62, U.S. Geol. Survey*, p. 47.
B. Quartz-monzonite, Mill Creek, Mont.—Lindgren, *Prof. Paper 27, U.S. Geol. Survey*, p. 18.
C. Quartz-monzonite, Hailey, Idaho.—Lindgren, *20th Ann. Rept., U.S. Geol. Survey*, p. 81.
D. Granite, Shafer Butte, Boise Co., Idaho.—Lindgren, *loc. cit.*
E. Biotite granite, Willow Creek Dist., Boise Co., Idaho.—Lindgren, *18th Ann. Rept., U.S. Geol. Survey*, p. 40.

The magmatic symbols for these rocks are as follows (calculation of A by Calkins, *loc. cit.*, and of C, D, and E by Washington, *Prof. Paper 14, U.S. Geol. Survey*):

A = I · 5 · 2 · 3, pulaskose.
B = I · 4 · 2 · 3, toscanose.
C = I · 4 · 2 · 3, toscanose.
D = I · 4 · 2 · 4, lassenose.
E = I · 4 · 3 · 4, yellowstonose.

[1] *Prof. Paper 62, U.S. Geol. Survey*, pp. 46–47.

The Cœur d'Alene rock, therefore, falls in a separate order, emphasizing its smaller quartz content. The striking difference in the habit of the ferromagnesian constituents is, of course, not shown by the quantitative classification. Rock of the Cœur d'Alene type is known also from the following localities:

1. On Vermillion Creek in Montana, where Calkins[1] found a rock "distinctly different from the intrusive masses to the west, and showing marked affinity with the masses of monzonite and syenite exposed to the southwest in the Cœur d'Alene district." The ferromagnesian mineral is either aegirine-augite or masses of hornblende and biotite pseudomorphic after pyroxene.

2. At the junction of Black Prince Creek with the St. Joe River where Pardee[2] found "a porphyritic monzonite" of "marked resemblance to the dominant rock in the intrusions of the Cœur d'Alene." Biotite, in addition to the distinctive hornblende, is reported to occur in this rock, but the quartz is in rather small amount.

3. On Gold Hill in the northern part of Latah County I have found a porphyritic, hornblendic monzonite that can in no way be distinguished from specimens from the Cœur d'Alene. Only twenty miles south of this locality are the Thatuna Hills, composed of the micaceous, quartzitic monzonite described as characteristic of the great batholith of central Idaho. In fact, it was the striking difference between the rocks from these two places that first called my attention to this problem. Unfortunately the intervening country is covered by Tertiary basalt, so the relationship between the rocks cannot be determined.

The other intrusives in northern Idaho, although showing considerable variation from the central Idaho type, nowhere, as far as I can learn, resemble this characteristic hornblende monzonite, and in most places they show marked affinities with the central Idaho type.

Reference to the map (Fig. 1) shows that the occurrences of the Cœur d'Alene type all fall in the northeast-southwest line that has already been noted by Calkins for the Cœur d'Alene and Vermillion

[1] *Bull. 384, U.S. Geol. Survey*, p. 47.

[2] *Bull. 470, U.S. Geol. Survey*, Pt. I, p. 46.

Creek outcrops. When, in addition to this striking linear arrangement, we consider the fact that nowhere throughout the other extensive areas of Idaho acidic intrusives as described by Lindgren,

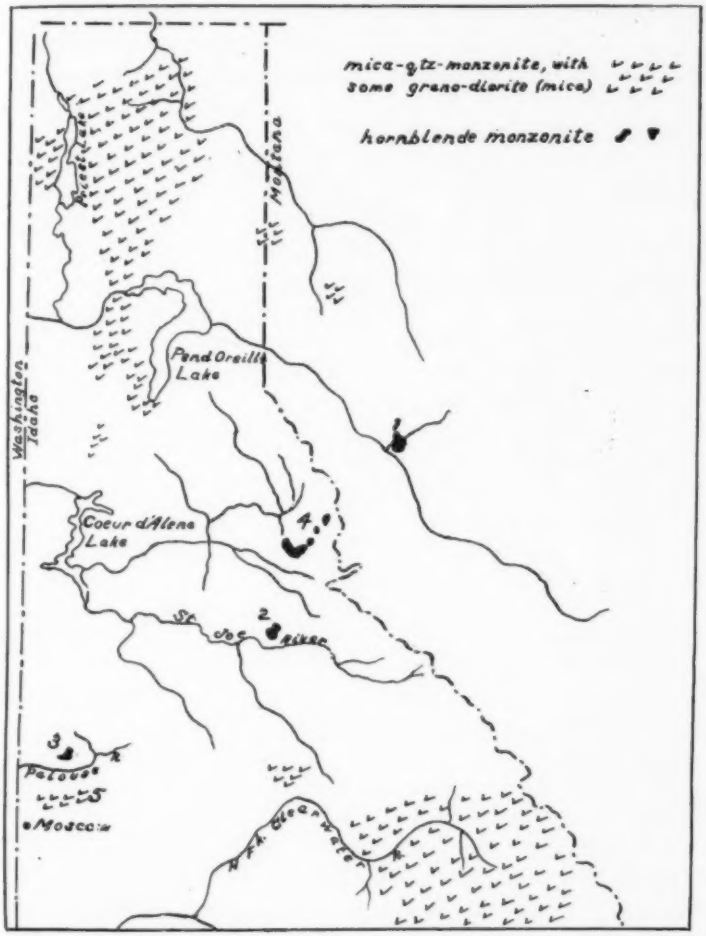

FIG. 1.—Acidic intrusives in Northern Idaho: 1, Vermillion Creek, Montana; 2, Black Prince Creek; 3, Gold Hill; 4, Cœur d'Alene District; 5, Thatuna Hills.

Calkins, Eldridge, and Umpleby, or where seen by Professor Livingston or the writer, is there any rock like this Cœur d'Alene type, it seems evident that we are dealing with something more significant than a single vast batholith underlying all of the state.

It is of course possible that the main Idaho batholith does underlie the Cœur d'Alene and neighboring districts at great depths, and that the exposures of the Cœur d'Alene type are offshoots reaching up into the overlying sediments and slightly differentiated. In view, however, of the proximity and nearly equal elevation of the Gold Hill and Thatuna Hills occurrences, I am inclined to regard the Cœur d'Alene type as the result of a separate intrusion, connected indeed in ultimate origin with the main batholith, but nevertheless distinct from it. Very probably it should be classed as a complementary intrusive, somewhat later than the main mass but from the same source. If this be true, we may expect to find monzonite of the Cœur d'Alene type cutting the micaceous quartz-monzonite. Until such a discovery is made, we must hold the problem in abeyance, content with the fact that the Cœur d'Alene intrusives, and those northeast and southwest of them, are distinctly different from the average type of Idaho quartz-monzonite.

A GRAPHIC METHOD OF REPRESENTING THE CHEMICAL RELATIONS OF A PETROGRAPHIC PROVINCE

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During recent years questions concerning the chemical composition of rock magmas, and the relation of these to one another, have been made the subject of extended investigation on the part of petrographers. A great number of rock types, and of many variations presented by these types, have been analyzed with great accuracy, so that now the material available for such studies has enormously increased.

As it is a matter of difficulty in comparing a long series of analyses to grasp clearly their various points of resemblance or difference, and to recognize the relationships, often more or less obscure, presented by the different types, investigators have from time to time sought to express the results of these analyses graphically so that by a comparative study of forms rather than figures the relationships in question might be brought out more clearly and presented in a more striking manner.

It is unnecessary here to consider in detail the various methods of graphic representation suggested by different workers. A review of the various methods proposed will be found in a paper by Iddings which appeared some years since.[1] Some of these methods, as for instance that of Reyer, show the relative proportions of the several chemical constituents present in the rock by means of a figure in which areas representing the various constituents are variously shaded or distinguished by different conventions. Others, such as that employed by Harker, indicate the relative proportions of the constituents by means of a curve on a plain surface. In a third class, such as those used by Brögger, the composition of the rock is represented by a geometrical outline whose shape would vary as the composition of the rock changed.

[1] J. P. Iddings, *Prof. Paper 18, U.S. Geol. Survey*, 1903.

Each of these methods of graphic representation has its peculiar merits, and each has also its disadvantages, which become manifest when it is desired by means of them to compare a long series of analyses, as for instance those representing a whole petrographic province.

In connection with some studies on the character and relations of the very striking series of alkaline rocks composing the petrographical province of the Monteregian Hills, which are now being carried on at McGill University, the attempt has been made to secure a more satisfactory expression of the chemical relations of the rocks of this province by employing a graphic representation in three dimensions. It is desired in the present brief paper to give an outline of the method employed and the result obtained.

Each of the analyses, 36 in number, was first plotted in the form of a curve which showed the actual and relative proportions of each of the chief constituents present in the rock. These are: silica, titanic acid, alumina, ferric oxide, ferrous oxide, lime, magnesia, potash, soda, water, carbonic dioxide.

The manganous oxide present was placed with the ferrous oxide and any small percentages of baryta or strontia with the lime. For purposes of simplicity other constituents, such as chlorine and sulphuric acid, which are occasionally present in small amount, were neglected, although of course it would be possible to represent them in the curve were it considered desirable to do so.

This curve was constructed by drawing a horizontal line, and marking off on it a series of points at equal distances from one another. At each of these points an ordinate was erected. On the first of the ordinates was plotted the molecular proportion of silica present in the rock. On the next ordinate the amount of titanic acid was similarly plotted, and on each of the others in succession the molecular proportions of the other constituents of the rock in the order enumerated above was shown. The points so obtained were then connected, and a curve thus constructed which shows in graphic form the chemical composition of the rock. In Fig. 1 there is seen the curve thus obtained for a camptonite occurring in the form of a dyke cutting the Trenton limestone at

the reservoir extension on the slope of Mt. Royal. A curve was obtained in a similar manner for each of the 36 analyses of the series.

These curves were then combined so as to give a value to the third dimension. For this purpose each curve was traced on a

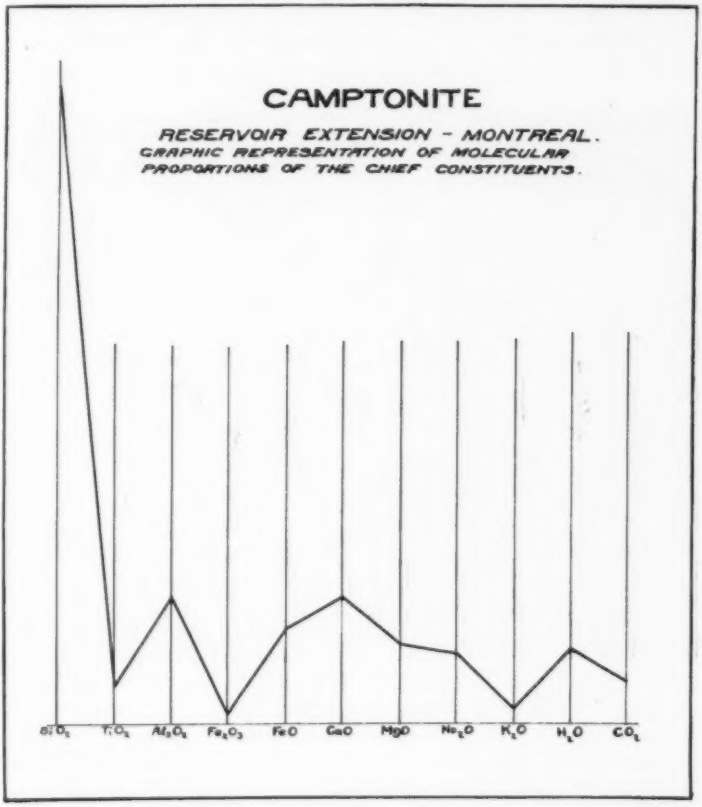

FIG. 1

thin rectangular sheet of metal, which was then cut so that its upper edge presented the outline of the curve, while the other three edges retained their rectilinear character. The form of the sheet representing the analysis of the camptonite has thus exactly the form of the diagram shown in Fig. 1.

These sheets of metal were then arranged in an upright position in a stout wooden frame, one in front of the other, at a distance of one inch apart. That of the rock having the highest content of silica was placed at one end of the series, while the others were arranged in the order of decreasing silica content, the most basic rock occupying the other end of the series. The spaces between the plates were then filled in with plaster of paris, the plaster

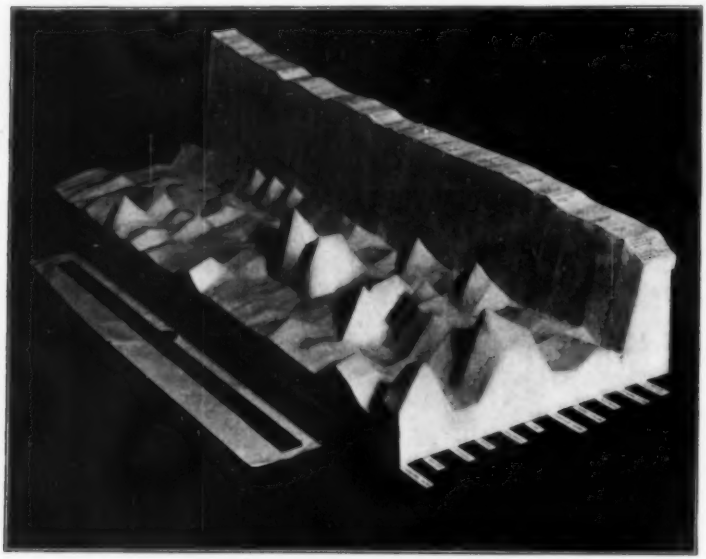

FIG. 2.—Model showing the chemical composition of the various rock types of the petrographical province of the Monteregian Hills.

between the successive sheets being smoothed down, so that the model thus completed presented a warped surface, passing transversely across which can be distinctly seen the traces of the curves representing the composition of the constituent rocks of the series.

The model thus obtained is shown in Fig. 2. A margin about one inch in width was left on the silica side of the model, to give space opposite the curve of each analysis to attach a small label having printed on it the name of the rock and the locality from

which it was obtained. In a similar manner the base on which the model rests is made to project a short distance at one end, so that another set of labels showing the chemical constituents represented in the analyses may be placed upon it. In this way, looking down the length of the model the character of the variation in the content of any chemical constituent in the series of rocks composing this petrographical province can be seen at a glance.

The predominance of any constituent, or group of constituents, in a certain part of the series is shown by a hill rising from the surface of the model, the shape of the hill varying according to minor variations in chemical composition of the rocks of this portion of the series. Depressions, on the other hand, indicate low percentages of a constituent, or group of constituents, in certain portions of the series.

The model shown in Fig. 2, representing the chemical relations of the rocks constituting the Monteregian Hills, combines the results of 36 analyses, the rocks ranging in composition from the acid nordmarkite of Mt. Shefford, containing 65.43 per cent of silica, at one end of the series, to the basic alnoite of Point St. Charles, near Montreal, holding 29.24 per cent of silica, at the other end. The intervening portion of the model shows the chemical composition and the mutual relations of the magmas of intervening acidity represented by the pulaskites, nepheline syenites, essexites, rouvellites, tinguaites, camptonites, montrealites, tawites, rougemontites, yamaskites, monchiquites, etc., of this petrographical province.

The model is 36 inches long, 12 inches wide, and $8\frac{1}{2}$ inches high at the highest part. It appears to present in a clear and rather striking manner the chemical relations of the alkaline magmas of this peculiar petrographical province, and similar models might be readily constructed which would set forth the characteristic relations of other provinces.

THE MODE OF FORMATION OF CERTAIN GNEISSES IN THE HIGHLANDS OF NEW JERSEY—*Concluded*

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PART II. DISCUSSION OF CERTAIN THEORETICAL PRINCIPLES INVOLVED IN THE INJECTION OF THE MAGMA

It has been pointed out that the granitic magma which was intruded into the original series seems to have been of a rather thin consistency. Nevertheless the degree of fluidity could hardly have approached, even approximately, that of water or an ordinary aqueous solution. The manner in which strata have been crumpled and twisted by movements of the magma argues a very effective degree of viscosity, as does also the support seen to have been offered to thinly tabular sheets of strata standing in a nearly upright position and evidently free for considerable distances from other support than that given by the adjacent magma. On the other hand, certain phenomena indicate that magmatic fluid was absorbed by the strata with little difficulty. At first sight these properties appear quite inconsistent with each other; nevertheless it seems possible to harmonize them. In order to derive a probable explanation of this and certain other phenomena it is necessary to consider certain of the properties of magmatic solutions.

At the present time it is well recognized that the temperature of fusion of the individual minerals of a rock, or even the temperatures required to fuse the aggregate, afford no indication of the temperatures at which the magma from which the rock crystallized was liquid in the interior of the earth. Since the time of Elie de Beaumont[1] the importance of the rôle played by water and other so-called mineralizers in lowering the temperature of fusion and increasing the liquidity of magmas has been perceived; and acidic magmas are regarded, for a number of reasons, as having

[1] Elie de Beaumont, "Note sur les émanations volcaniques et métallifères," *Bull. soc. géol. fran.* (2), IV (1847), 1249–1333.

contained large amounts of water. Nevertheless the property of critical temperatures of liquids appears at times to be somewhat of an obstacle to a clear comprehension of the miscibility of rock magma with water and other volatile substances at a high temperature. The critical phenomena seem sometimes to be understood to imply that at temperatures above 374° C. water, under all circumstances, ceases to be a liquid and becomes a gas. A better statement would be that at the critical temperature and at the corresponding pressure the properties of liquid and vapor cease to differ. The discontinuity of properties which characterizes the transformation from liquid to vapor at lower temperatures and pressures disappears. Moreover, the critical temperature of a liquid is not a fixed point but varies with the amount of material in solution. With pure water the critical phenomena appear at 374° C., but when water holds material in solution the vapor pressure is lowered and the critical point is raised. The greater the amount of dissolved material the greater is the displacement.

These facts have been emphasized in an investigation recently carried out by G. W. Morey[1] in the Geophysical Laboratory, of which the results have just been published. In one experiment 2 gm. of a glass of the composition $K_2O:1.7\ SiO_2$ was heated with 4.842 gm. water to a temperature of about 360° in a gold crucible within a gas-tight steel bomb. The result was a pasty glass containing 32 per cent water. In another experiment 2 gm. of glass of the same composition as before and 5 gm. water were exposed to a temperature of 490° under the same conditions. The product obtained was a hard glass containing 20 per cent water.[2] In the latter experiment the temperature was far below that at which the dry materials would begin to melt and at the same time more than 100° above the critical temperature of water alone, and the formation of a strongly hydrated glass is a striking phenomenon.

There is, therefore, no theoretical difficulty in supposing that fused silicates may form a homogeneous solution with water and other volatile substances at high temperatures, and the properties

[1] G. W. Morey, *Jour. Amer. Chem. Soc.*, XXXVI, 2 (1914), 215–30.

[2] Evidence was obtained which showed that even those glasses which became highly rigid when cooled formed mobile liquids at the temperature of the experiments.

imparted by the presence of such volatile matter—increase of mobility and greater chemical activity—will be retained.

The ease of flow manifested by the New Jersey granite which has been described is therefore quite understandable; nevertheless it seems to have retained a quite effectual degree of viscosity, as has been pointed out in describing the field phenomena. The facility, then, with which magmatic material seems to have been transfused into the adjacent strata is not at once intelligible. The difficulties in harmonizing the phenomena which are exhibited will be better appreciated by a calculation of the rate at which a fluid under pressure enters minute pores. For instance, an approximate calculation by Poiseuille's formula shows that if we assume a liquid having a coefficient of viscosity of 1.0 (about one-tenth that of glycerine at 19° C. or 100 times greater than that of water at the same temperature) and conceive it to be pressed with a force of 810,000 gm. per square cm. (corresponding to an overhead load of about 3 km. of rock strata) against a surface of 1 sq. cm. pierced with 1,000 holes, each having a radius of 0.0001 cm., and suppose the thickness of the partition to be 1 dm., then in one year only 0.1 c.c. would ooze out on the far side.

The assumptions made are far from exact. The form and disposition of the pores in rocks do not correspond to the straight cylindrical tubes to which Poiseuille's formula closely applies, and the exact degree of viscosity is unknown, but the most important factor is the diameter of the pores, for the flow is about as the fourth power of this dimension. Thus if we had assumed pores of a radius of 0.0004 cm. (probably an excessive estimate of size for the pores of ordinary rocks under heavy load) we should have derived a value of 25 c.c. per year.

It seems, on the whole, that we can expect little movement of the magma as a unit into minute pores, and we shall have to seek an explanation of its transfusion into the adjacent rock from other considerations.

Among the phenomena shown in the field several facts merit emphasis. As I have already mentioned, the introduction of the magma seems to have been effected with remarkably little disturbance of the previously existent relations of the original layers.

It is true that the larger bodies of granite show a tendency toward a lenticular outline, indicating that in such instances the adjacent rock has been forced apart; and it is found also that flexures of a minor order have been produced in the layers, but the striking fact remains that frequently evidences of the original structure survive in the midst of large masses of granite and show little disturbance of position. The most noticeable evidence in this respect is the parallelism shown by the bands of inclusions with the regional strike of the gneisses. This is observable even in those cases where assimilation by the magma has proceeded so far that the included material no longer appears as distinct bands but merely as dark-colored *schlieren* of indefinite outline. Moreover, the presence of faint parallel lines of dark minerals within the granite alongside of inclusions, as well as a certain parallelism visible in the arrangement of the component minerals of granitic masses in which no distinct xenoliths are perceptible, implies that some sort of structural framework of solid material was retained. It cannot be supposed, therefore, that the injection of the magma was of a sudden and violent nature or that it entered by simply wedging the original rock apart along the dividing laminae and occupying the spaces made. On the contrary, it must have gained access in such a slow and quiet manner that in many places delicate structural relations were left undisturbed. In fact the process appears to have possessed many of the features of a gradual substitution rather than the violent characteristics of intrusion as we often picture them.

The advocates of *lit-par-lit* injection to whom reference has been made have supported the theory that the advance of the main body of magma was preceded by that of a more attenuated portion. They believe that these solutions in advance, which were probably similar in composition to that of the main body but characterized by the presence of a larger percentage of the volatile ingredients, exercised functions of a preparatory character, metamorphosing the material into which they penetrated and rendering it gradually more susceptible to the action of the magma which followed. Lacroix's conception[1] may be illustrated by the following quotation,

[1] A. Lacroix, "Le granite des Pyrénées et ses phénomènes de contact," *Carte géol. de France, No. 71* (1900), p. 26.

in which he describes certain contact rocks formed by the action of a granite batholith upon schists:

> The characteristic of the contact phenomena of this type consists essentially in the addition of volatile or transportable materials, emanating from the magma and modifying generally in a manner more or less profound the chemical and mineralogical constitution of the sediments traversed. . . . .
>
> In the regions which form the object of this study, and in many others, the disposition of the sedimentary beds in relation to the granite, the existence in the midst of the latter of sedimentary fragments which, in spite of their transformation, have preserved the same orientation as the peripheral schists and limestones to which they belong, finally the absence of important dislocations at the immediate contacts, tend to show that the granitic magma has been brought to its position in a slow and progressive fashion by imbibition and dissolution of the sediments for which it has been substituted.
>
> In proportion as the granitic magma appeared in the sedimentary beds it was preceded and accompanied there by its cortège of transforming emanations, and what it had to dissolve to make place for itself was no longer normal sediments, but transfused rocks, transformed or in process of transformation by fixation of the emanations which had gone out from its own mass.
>
> The most intense modifications undergone by the schists metamorphosed into leptynolites consist essentially in the development of a large quantity of feldspars, quartz, biotite, etc. The limit toward which these leptynolites, which are often comparable from a mineralogical standpoint with gneisses, tend is the mineralogical composition of granite itself.

In a recent article Sederholm[1] describes contact-effects of a rapakivi granite upon older hornblende-schists, with introduction of typical minerals from the rapakivi. He is able to trace out the transitional changes in the minerals of the schist by absorption of material from the granite. "Microscopically as well as macroscopically it is shown, therefore, that the whole schist-mass has been permeable for gases and juices (*Säfte*) which have gone out of the granite magma, as well as for the magma itself in further progress of change. This contact would have convinced even the most zealous anti-injectionists."

The ideas which are formulated in these quotations have met with opposition from some geologists, but the objections seem to be based partly upon a misconception. The advocates of the

[1] J. J. Sederholm, "Ueber ptygmatische Faltungen," *Neu. Jb.*, XXXVI (September, 1913), Beilage-Band, p. 491.

doctrines have evidently had in mind their application to a certain type of phenomenon and do not imply that all granitic intrusions have occurred under like conditions or been accompanied by similar effects. It seems to be considered at times that the observations of Lacroix in the Pyrenees are opposed by those of Brögger in the Christiania region, but these two geologists themselves evidently perceived that they had to do with processes of a different nature and that the same phenomena should not be sought or the same explanation applied. Lacroix[1] has discussed Brögger's views and says: "None of the arguments which this scientist deduces from the study of his region bears against the theory of assimilation, applied to the Haute-Ariège; the facts which he has described are in effect the antithesis of those which I have set forth in this memoir." He suggests that the dissimilarities may be attributed to differences in the depth at which the intrusions took place in the two regions.

Brögger[2] also states that it would not be justifiable without further evidence to apply his observations on the Christiania district to the granite regions of the older primitive rocks and regionally metamorphosed fold-mountains. Evidently both Lacroix and Brögger recognize the fact that the type of granitic intrusion to which each has devoted his attention is not the only form which is found.

The relations which Michel-Lévy, Lacroix, Sederholm, and others have described offer an array of evidence in confirmation of their views, and there seems to be no inherent obstacle to accepting them.

In trying to account for the formation of a dilute magmatic solution in advance of the main invasion several possibilities suggest themselves.

It appears probable that the escape of volatile constituents through pores of the wall-rock too minute in size to permit unchanged magma to follow would have a tendency in this direction. Field evidence appears to warrant the view that considerable quantities of vapors have often escaped in this manner and that

[1] A. Lacroix, "Le granite des Pyrénées et ses phénomènes de contact," *Carte géol. de France, No. 64* (1898), p. 63.

[2] W. C. Brögger, *Die Eruptivgesteine des Christiania-Gebietes*, II (1895), 152.

important results have been accomplished through their agency. We must suppose that these gases will exert a fluxing or solvent action upon the minerals of the walls, so that some fraction of them shall form a solution with these minerals, giving rise to a sort of secondary magma within the walls.

Moreover, during the advance of streams of magma between layers of rock, such as is characteristic of *lit-par-lit* injection, it seems almost inevitable that by contact with the cooler walls a portion of their load of dissolved material should be deposited. Thus the solutions ahead would become progressively more dilute, but with the rise of temperature ultimately produced in the walls by the stores of heat imparted to them by a number of closely adjacent streams of magmatic solution unchanged magma would finally enter among the layers and carry farther toward a conclusion the processes of transformation initiated by the solutions in advance.

By whatever means a differentiation of the magma-streams is effected, the advance of the magma into the wall-rock would be attended by phenomena of various sorts, such as impregnation of the walls, solution and removal of some of the components of the wall-rock, and reactions with the minerals with which the solution came into contact.

When the chemical nature of the wall-rock differs greatly from that of the magma the reactions between the two might be expected to effect striking results. Limestones seem especially fitted to react with the magma, but other rocks may participate in a like manner. For this reason it may not be possible to decide what the nature of the original rock has been, as was pointed out in considering the character of the strata invaded by the magma in the New Jersey example. Instances of the extreme effect produced by such reactions were found by Adams and Barlow[1] in the Haliburton-Bancroft area. A passage in their publication refers to a certain contact of a granite batholith with the limestone wall-rock, and reads as follows:

> . . . . The limestone bands fade away imperceptibly into the amphibolite, the latter being undoubtedly produced by the alteration of the limestone. These rocks are invaded by the granite, traversing them in apophyses which swarm

[1] Adams and Barlow, "Geology of the Haliburton and Bancroft Areas," *Memoir No. 6, Canadian Geological Survey* (1910), p. 101.

through them in all directions, often running parallel to the banding, and elsewhere cutting across it. . . . . The alteration is due not only to the proximity of the main mass of the batholith, but to the immense amount of granitic material which occurs intruded through the series, sometimes in large masses, but very frequently in thin bands which have found their way in between the beds of the invaded limestone, changing it into amphibolite, and presenting a typical instance of *lit-par-lit* injection. The granite, furthermore, not only penetrates this amphibolite series, but floats off masses of it, which, in the form of bands, streaks, and isolated shreds, are seen thickly scattered through the granite in the vicinity of the contact. . . . . The separate fragments of amphibolite, where completely surrounded by the granite, while clearly nothing more than masses of altered limestone, are rather harder, and have a more granitized appearance than the rock which is still interstratified with the limestone.

From the various descriptions which have been cited it appears that the type of intrusion to which the term *lit-par-lit* injection has been applied presents evidences that the invasion of the magma is preceded by the advance of a wave of metamorphism into the wall-rock, by which the character and composition of the original material are radically altered. By the deposition of magmatic minerals and by the removal in solution of certain of the previous constituents, the composition tends to approach that of the magma itself, and when blocks of wall-rock are finally engulfed in the magma their composition may be so changed that their assimilation effects but little change in the composition of the latter.

If this conception is well-grounded the zone of metamorphism which surrounds areas of *lit-par-lit* injection is not to be considered as wholly an after-effect of intrusion, due to expulsion of volatile substances during the consolidation of the magmatic mass, but also as due to a preliminary process, ultimately leading up to an invasion of the magma and assimilation of the altered material.

SUMMARY

A description has been given of the structural relations observed in a certain area of banded gneisses of pre-Cambrian age in Northern New Jersey, where unusually favorable conditions for observation have been found. The structures shown here are believed to be typical of those prevailing over considerable areas in this portion of the state. Evidence is given leading to the belief that the structures at this locality cannot well be attributed to the squeezing-out

of a partly differentiated magma or to the shearing and recrystallization of a solidified rock, but that their origin must be looked for in a process involving the injection of a thinly fluid granitic magma between the layers of an original rock of laminated structure. Structures which still survive in the larger bodies of granite indicate that the process of injection was carried out in a most quiet and gradual manner, and possessed many of the characteristics of a substitution of the original material by the magmatic solution rather than the features of a violent intrusion. The observed relations are very similar to those which French geologists have described under the name of *lit-par-lit* injection, and the mode of operation is believed to have been essentially the same.

Certain features observed in the gneisses imply properties of the magma which at first sight do not appear mutually consistent. Thus the degree of viscosity indicated by the presence of thinly tabular sheets of inclusions within the granite, standing nearly upright and unsupported except by the magma on either side, does not harmonize with the facility with which magmatic material has been transfused into the original rock. In trying to reconcile these features inquiry has been directed toward a consideration of certain of the properties of magmatic solutions. The question of the critical temperatures of volatile substances is discussed in its bearing upon their condition within the magma. Further, the problem of a possible differentiation of a magma when injected into a wall-rock in a multitude of adjacent streams is taken up, as related to the views expressed by the advocates of *lit-par-lit* injection. Several methods by which this might be accomplished appear possible, and it is suggested that the escape of gases from the magma into the wall-rock would have a tendency in this direction and that at the same time the deposition of magmatic minerals by contact of tongues of magma with the cooler walls would effect results of a similar nature. Thus the advance of the main body of magma would be preceded by that of a more dilute portion, which would be able to impregnate the wall-rock with facility and initiate processes of transformation and solution which the more concentrated body following would carry farther toward completion.

December, 1913

A COMPARISON OF THE CAMBRIAN AND ORDOVICIAN RIPPLE-MARKS FOUND AT OTTAWA, CANADA[1]

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Introduction.—We are told by an eminent geographer that "the most important part of the geological labors of the ocean is hidden from our eyes."[2] The evident truth of this observation affords a sufficient reason for making a careful record of every fact which may help to throw light on the processes of sedimentation. Ripple-marks on sandstones and sandy shales are among the most familiar of the characters with which marine sediments have been impressed, but we have yet much to learn concerning them. These features when fully understood will tell us much more of the local physical conditions under which they were formed than we are at present able to infer from them. It still remains for future research to determine whether a law can be deduced which will connect the length and height of waves with the width of the ripple-marks resulting from the simultaneous water oscillation. Among the present needs in connection with the effort to acquire a more fundamental knowledge of ripple-mark phenomena appear to be recorded data on (1) the different types of ripple-marks to be met with in a given set of beds where the factors of depth and texture of sediment are uniform, (2) characteristics which distinguish ripple-marks associated with particular kinds of sediment from those found in other sediments.

The observations in this paper are recorded with this twofold object in mind. The Cambrian and Ordovician sediments both show well-developed ripple-marks in horizons which are exposed in the Ottawa district. The Cambrian ripple-marks which will be described occur in sandstone, while those in the Ordovician are impressed upon limestone.

[1] Published with the permission of the Director of the Geological Survey of Canada.

[2] Elisée Reclus, *A New Physical Geography*, Vol. II (1890), "The Ocean," p. 103.

Cambrian ripple-marks.—Cambrian sandstone is extensively exposed in a flat-topped hill 12 miles west of Ottawa, just south of the intersection of the Canadian Northern and Grand Trunk railways. The rock is a hard, white to buffish-gray, moderately coarse sandstone. It is thin-bedded, lying in strata 2–10 inches thick. Some of the strata are beautifully ripple-marked.[1] All of the ripple-marked beds and those associated with them appear to be quite barren of fossils, although a *Lingulepis acuminata* Con. occurs abundantly a little lower in the section along the railroad. Two

FIG. 1.—Ripple-marks in Cambrian sandstone. The large slab shows the symmetrical type with low ridges and broad flat troughs; the asymmetrical type is represented by the small slab at the right of the picture.

distinct types of ripple-marks occur in the highest beds exposed on the hilltop. Both of these are shown in Fig. 1. The type seen in the smaller slab on the right of the picture is better shown in Fig. 2. The ripple-marks shown on the large slab are symmetrical, the two sides of the ripple ridge having the same slope or curvature. Those seen on the smaller slab and in Fig. 2 are asymmetrical, the slope of the ridges on one side being much steeper than that of the ridges on the other side. The symmetrical ripple-marks show some features in which they differ from those usually met with in sandstones.

[1] I am indebted to Mr. L. D. Burling for directing my attention to these ripple-marks.

Instead of the usual rounded or angular trough between the crests, the ripple ridges rise from interspaces which are almost or quite flat. The gently rounded parallel ridges are separated by flat interspaces generally one and one-half to two times their width. Traces of a diminutive ridge in the middle of the flat-bottomed trough may sometimes be noted. This type of ripple-mark is comparable in its essential features with, though not precisely like, a form of ripple mark which I have seen developed in Lake Deschênes near Ottawa. A photograph of these sand ripple-marks made from a cast of a mold taken under water at Lake Deschênes is here reproduced in Fig. 3. Comparison of the two figures shows broad, nearly flat troughs in each separating the comparatively narrow rounded ridges. In the Lake Deschênes specimens the miniature secondary ripple-marks are more clearly developed than in the sandstone ripple-marks. The lake ripple-marks were formed under natural conditions in water 6 inches deep and were under observation during and after their development for a period of three days. They are the product of water oscillation resulting from a very gentle on-shore breeze. This breeze and the accompanying wavelets were not strong enough to produce the asymmetrical type of ripple-mark with steep leeward and gentle windward slopes which results from more vigorous movement of the water and leaves a record of the direction of the movement of

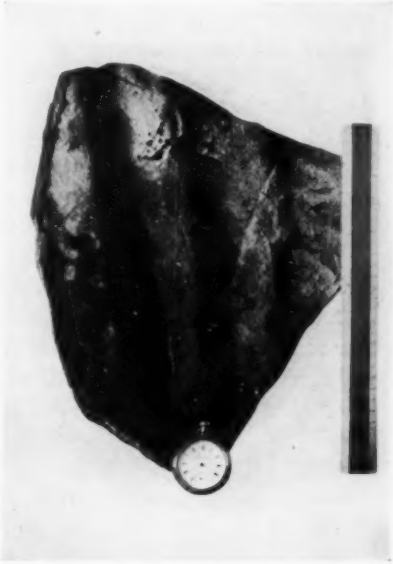

FIG. 2.—Cambrian sandstone slab showing asymmetrical type of ripple-mark.

the waves or currents concerned in its production. The similarity in type of the Cambrian and Lake Deschênes ripple-marks shown in Figs. 1 and 2 suggests the inference that the Cambrian impressions probably represent the work of winds and waves of a very gentle character.

In beds which lie within a very few inches of the symmetrical ripple-marks which have been described occur the asymmetrical ripple-mark type shown in Fig. 2. The ripple-marks in the specimen shown in Fig. 2 have a width from crest to crest of 3 inches and a trough depth of one-fourth inch. Their lee and stoss slopes are, respectively, 9°–15° and 20°. These ripple-marks trend about S. 35 E. and present their steep sides toward the northeast. Ripple-marks of this type represent either a wind or water current moving in a northeasterly direction. An interesting feature of these ripple-marks is the presence of a second set of much finer and less distinct ripple-marks which show sometimes clearly but generally faintly on the stoss slopes of the prominent ripple-marks. These secondary ripple-marks cross the others at an angle of about 45°. If they are the product of wind rather than current action, they indicate a shift of wind of nearly a quadrant from the position which it must have maintained during the development of the larger ripple-marks. At the time these Cambrian sands were deposited the nearest shore of the Cambrian sea lay five or six miles to the northeast if we may judge from the close proximity and relative elevation of the archean rocks in the Laurentian mountains on the opposite side of the Ottawa valley. The wind or current direction indicated by both of these sets of ripple-marks therefore appears to indicate a general on-shore direction.

FIG. 3.—Cast from mold of sand ripple-marks taken under water at Aylmer, Quebec. Compare with Fig. 1.

Trenton limestone ripple-marks.—The extensive quarry in the Trenton limestone at the cement mill one-half mile northwest of Hull, Quebec, exposes a section with a vertical thickness of nearly 100 feet. Three or four horizons in this section show, in the vertical walls of the quarry face, contacts between adjacent beds which are very suggestive of ripple-marks of large dimensions. The imperfect exposures of the lower of these horizons leaves room for some question as to their true nature but it appears clear that the uppermost is an example of ripple-marks. Only the uppermost of these horizons, which is admirably exposed for study through the stripping of the stone to this level over an area of considerable extent, will be considered here.[1] This bed lies 10–15 feet below the top of the quarry in the part of the section called by Dr. Percy Raymond[2] the crinoid beds.

The limestone here is a dark-gray rock lying in strata usually 6 inches to a foot in thickness. The regular horizontal contact of these strata gives way at the horizon under discussion to a billowy trough and crest contact. This is much less regular in profile than the profile of the ordinary ripple-mark as seen in most sandstones. The somewhat irregularly rounded summit of the ridges and their variable width might suggest on brief examination an unconformity; their parallelism and the identity of the fauna on both sides of this horizon, however, render this explanation untenable. The ridges trend approximately east and west and their axes are from 2 to 3 feet apart. No sharp crests are to be seen. All of them, as shown in the photographs (Figs. 4 and 5), are rounded or somewhat depressed on top. The troughs have an average depth of about 6 inches below the crest and frequently coalesce, thereby producing many short ridges. An interesting feature connected with these ripple-marks is the occurrence on them at some points of numerous crinoid stems some of which have the heads attached as shown in Fig. 6. These are often two feet or more in length and indicate by their presence that the crinoids lived on this part of the sea bottom shortly after the formation of the ripple-ridges in water

[1] These were brought to my notice by Mr. W. A. Johnson.

[2] "Excursions in Neighborhood of Montreal and Ottawa," *Guide Book No. 3* (1913), p. 143.

which was sufficiently quiet not to break up the stems into the small sections usually found in limestone. Above the ripple-marks 6 or 8 inches, however, there is equally clear evidence of current action

FIG. 4.—General view of ripple-marks in Trenton limestone at cement quarry, Hull, Quebec.

FIG. 5.—Close view of ripple-marks in quarry at Hull, Quebec

at a later period having been involved in the distribution of the materials comprising some of the beds. At this level a band of cross-bedded limestone a few inches in thickness occurs with layers inclined at about 27°. The material comprising these current-

built beds consists largely of small fragments of crinoid stems and other fossils.

Comparison of limestone and sandstone ripple-marks.—These Trenton limestone ripple-marks show a striking contrast with those of the Cambrian sandstone which have been described, in the matter of size. Instead of an amplitude of 2 or 3 inches which characterizes the Cambrian ripple-marks, these have a width of 2 or 3 feet from crest to crest. This discrepancy in size is of interest because it exists between the great majority of the examples of limestone ripple-marks and sandstone ripple-marks which have come under my notice. I have observed very few cases where the amplitude of ripple-marks in sandstone exceeded a few inches.[1] On the other hand, six of the eight photographs in my collection illustrating limestone ripple-marks, which represent as many different geological horizons and widely separated localities, show an amplitude of more than one foot. The data given in geological literature relative to limestone ripple-marks are often too incomplete to furnish data on the amplitude, but where given the amplitude is likely to be represented by a figure much larger than those which usually represent the amplitude of sandstone ripple-marks.

The earliest description of limestone ripple-marks with which I am acquainted relates to ripple-marks of large amplitude. This occurs in Dr. John Lock's description[2] of the "Blue Limestone" (Ordovician) of southwestern Ohio. Under the head of "Waved Strata," he described a stratum in which "the upper side is fluted out in long troughs 2–3 feet wide and about 2 or 3 inches deep, the edge or ridge between them being generally sharp. . . . . These waves are not local, but may be traced in the same stratum over tracts of many miles."[3]

Dr. August Foerste states[4] that "wave-marks (ripple-marks) occur in Ohio, Indiana, and Kentucky, in abundance in the Lower Eden, Upper Richmond, and upper Brassfield limestones. They

[1] Mr. G. K. Gilbert has recorded one exceptional case in the Medina sandstone in which the amplitude is reported to be several feet (*Bull. Geol. Soc. Am.*, X, 135).

[2] *Geol. Surv. of Ohio* (1838), p. 246, Pl. 6.

[3] Lock expressed the belief that these were not ripple-marks because "all geologists will agree that the blue limestone has been formed far below the reach of ripples."

[4] Letter to the writer.

occur in great numbers, but not so abundant, also in the Middle Eden. In Kentucky they are common also locally in the Mount Hope bed, at the base of the Maysville. They occur often near the middle of the Arnheim and at various intervals in the Lower and Middle Richmond in the three states mentioned." Dr. Foerste has recorded[1] a considerable number of observations on the limestone ripple-marks in the Ordovician and Silurian formations of these three states. Most of those which he observed appear to be of large size, those seen at one locality having crests 49 inches apart which rise $3\frac{1}{2}$ inches above the troughs.[2] Foerste remarks that "their unusual size will, however, attract attention even from one familiar with the work of the sea." Moore and Allen[3] have published photographs and descriptions of Upper Ordovician (Richmond) limestone ripple-marks with an average amplitude of 2.63 feet which occur in eastern Indiana.

In New York state, Miller[4] has described ripple-marks in the Trenton limestone of Lewis county which measure 24–56 inches from crest to crest with troughs 4–7 inches deep. At another locality in the same state, Professor Cushing[5] has described ripple-marks in the Trenton limestone with an amplitude of 9–15 inches and a depth of 1–3 inches. Cushing remarks concerning these that "they are considerably broader than the usual ripple-marks in sandstones." This observation, it may be noted, tends to confirm the opinion already expressed by me that sandstone ripple-marks have generally a comparatively small amplitude.

But few references to ripple-marks on Silurian limestones have come under my notice. Two of these which relate to New York and Ohio localities may be cited. East of Lockport, New York, Kindle and Taylor[6] report ripple-marks in the uppermost member

[1] "On Clinton Conglomerates and Wave-Marks in Ohio and Kentucky," *Jour. Geol.* III (1895); 1–10 (reprint). "The Richmond Group along the Western Side of the Cincinnati Anticline in Indiana and Kentucky," *Am. Geol.*, XXXI (1903), 333–61.

[2] *Jour. Geol.*, III, 37.

[3] *Proc. Ind. Acad. Sci. 1901* (1902), pp. 216–18, Pls. 1–3.

[4] *Bull. N.Y. State Mus. No. 135* (1910), p. 36.

[5] *Bull. N.Y. State Mus. Nat. Hist. No. 77* (1905), p. 34.

[6] *Folio U.S.G.S. No. 190* (1913), p. 7.

of the Clinton limestone which have crests 18–30 inches apart with troughs 2–4 inches deep. Foerste[1] mentions ripple-marks in the Springfield dolomite. west of Peebles, Ohio, one-fourth mile. Professor Prosser[2] states that the ripple-marks at this locality occur in two beds. "In the lower layer the crests of the ripples are 22 inches apart and the trough 4 inches deep. In the upper layer the crests are 45 inches apart and the troughs 4½ inches deep."

In the Devonian, I am acquainted with but three examples[3] of ripple-marked limestones. Each of these shows an amplitude of 2 feet or more.

The only notice of Mesozoic limestone ripple-marks[4] which has come to my attention records the occurrence of ripple-marks in limestone of Jurassic age in Utah with an amplitude of 6–12 inches.

The examples cited above from the literature of limestone ripple-marks plainly indicate the large size usually attained by ripple-marks in pure limestone. It must be noted, however, that ripple-marks of small size also sometimes occur in limestone. A photograph of ripple-marks on a dolomitic limestone with a smaller amplitude than the sandstone ripple-marks reproduced in this paper has recently been published;[5] and Miller[6] mentions Ordovician examples with an amplitude of 1–2 inches. Foerste[7] has observed that large and small limestone ripple-marks in the Ordovician of the Ohio valley are seldom associated but confined to distinct horizons. One must therefore conclude that the generally larger size of limestone as compared with sandstone ripple-marks results from some factor other than the different physical characteristics of the two kinds of rock. It seems probable that the explanation is to be sought in the fact that limestones are generally formed somewhat farther from the shore and in deeper water than sandstones. The different average amplitude of sandstone and limestone

[1] *Jour. Cin. Soc. Nat. Hist.*, XVIII (1896), 167.

[2] Letter to the writer, February 6, 1914.

[3] One of these has been described (*Ottawa Nat.*, XXVI [1912], 1–3, Pl. 7).

[4] G. K. Gilbert, *Science*, III (1884), 375–76.

[5] Kindle and Taylor, *Folio U.S.G.S. No. 190* (1913), Pl. 25.

[6] *Bull. N.Y. State Mus. No. 135* (1910), p. 36.

[7] Letter to the writer.

ripple-marks may be related to the relative average depths at which limestone and sandstone sediments accumulate. In other words, the usual large amplitude of limestone ripple-marks may be a function related to the depth of the water in which they are formed. If we apply this as a working hypothesis to the case under consideration, we must conclude that the ripple-marks in the Trenton limestone near Ottawa were formed under water of considerably greater depth than those in the Cambrian sandstone. The presence of long

FIG. 6.—View showing profusion of crinoidal remains found on the surface of some of the ripple-marks in cement quarry at Hull. Photograph by L. D. Burling.

crinoid stems upon the limestone ripple-marks is significant in connection with the question of the depth of the water under which they were formed. They were developed during an interval when a luxuriant crinoid fauna flourished in this part of the Trenton sea. All of the members of this order which live in the present seas are found, with a very few exceptions, in comparatively deep water. The living stalked crinoids are found chiefly at depths of more than 200 fathoms; a few extend into 140 fathoms and a very limited number into 58 fathoms, while one or two species are known in still shallower water. Geologists are accustomed to infer the

former existence of shallow seas where corals are the dominant fossils, and it seems an equally safe deduction to infer a sea of considerable depth where stalked crinoids comprise the dominant element in the fossil fauna as they do in the ripple-mark horizon under consideration. The occurrence on the ripple-marks of numerous crinoids in which the head remains joined to nearly complete stems (Fig. 6) affords direct evidence that they lived in water of sufficient depth to be seldom disturbed by wave action. The well-known fragile character of the union of the individual segments of a crinoid column and of the head and column could not be expected to survive the bottom disturbance of ordinary wave action. It is probable that the ripple-marks on which the crinoids rest are the result of storm waves of unusual size whose oscillations penetrated to a depth not affected by ordinary waves. It appears from these considerations that both the physical and the faunal evidence suggest for the ripple-marks in the Trenton limestone an origin in water of greater depth than the ripple-marks in the Cambrian limestone.

NEW AND LITTLE-KNOWN INSECTS FROM THE MIOCENE OF FLORISSANT, COLORADO

T. D. A. COCKERELL
University of Colorado

TRICHOPTERA

Phryganea wickhami n.sp.

♂ Upper wing as preserved uniform light reddish brown, with the venation distinct but not dark; length 22 mm., width $8\frac{1}{2}$; apex of wing to base of discoidal cell $13\frac{3}{4}$ mm.; length of discoidal cell 6 mm., of cellula thyridii not quite $7\frac{1}{2}$; venation normal for a male, the media with only three branches; R_1 apically with a very strong double curve. The shape of the wing is practically as in *P. latissima* Ulmer, from amber; the venation differs from that of the amber species as follows: R_2 leaving discoidal cell more basad, about $2\frac{1}{3}$ mm. before R_3 leaves it; separation of M_1 from M_2 at same level as separation of R_3 from discoidal cell; upper apical corner of cellula thyridii (separation of M_2 from M_3) only a short distance basad of separation of R_2 from discoidal cell; oblique basal side of cell between Cu_1 and Cu_2 longer; cell below Cu_2 smaller. In the characters of venation wherein *P. wickhami* differs from *P. latissima*, it agrees closely with *P. singularis* Ulmer, another amber species.

From the other Florissant species of *Phryganea*, *P. wickhami* is distinguished especially by the shape of the wings, which are not elongated as in *P. miocenica* Ckll., or with the apical margin truncate as in *P. labefacta* Scudder.

Miocene shales of Florissant, Wilson Ranch (H. F. Wickham). At the same place Professor Wickham found one of each sex of *P. labefacta*.

NEUROPTERA

Raphidia pulveris n.sp.

Anterior wing 11 mm. long, not quite $3\frac{1}{3}$ broad; costal area broad, with ten cross-veins; stigma about 2 mm. long, the lower side a little over 1 mm., oblique cross-vein very distinct; two cells on costa beyond stigma; subcosta ending nearly 1 mm. below base of

stigma; three discoidal and three cubital cells; no closed cells beyond the discoidals; upper discoidal beginning at about level of end of subcosta, and ending a short distance beyond middle of lower side of stigma; second discoidal more produced basally than in *R. rhodopica*, its base not invaded by fork R_2–R_3, which falls some distance beyond its end; cell in fork R–R_5 scarcely hexagonal, its lower median face very short; R_2 branched at level of end of lower side of stigma, its upper branch branching again with each of the branches forked near margin, its lower branch simple; R_3 forked close to wing-margin; R_4 simple at end, but R_5 forked; six forks on lower margin of wing, not counting R_3 or R_5; cells below Cu_1 (not the cubital cells of descriptions) nearly as in *R. rhodopica*, except that the anal cells are different in detail, two large ones joined by a very narrow isthmus.

Posterior wing about 10 mm. long; apical field practically as in anterior wing, base and costa differing as usual. The end of R_2 is different, the second division of the upper branch being simpler while the lower branch has a very long narrow fork; R_3 is simple at end; there are six marginal forks basad of R_5, as in the upper wing, but the first is so long that its corner joins the end of the third discoidal cell; an almost rectangular cross is formed near the middle of the wing where the cross-veins leave the media above and below.

Miocene shales of Florissant, Wilson Ranch (H. F. Wickham). The upper wing is the type; the lower wing is on a different slab, but, from its close resemblance to the upper, evidently belongs to the same species. In Rohwer's table in *Am. Jour. Sci.*, XXVIII, 534, this runs to *R. mortua* Rohwer, but differs by the much larger number of cross-veins in the costal area, subcosta joining costa nearer stigma, and marginal V-shaped cells much more numerous. It is no doubt allied to *R. mortua* and *R. exhumata* Ckll., but there are so many small differences that I can only consider it distinct. There is also much resemblance to *Archiraphidia tumulata* (Scudd.), but there are too many differences to regard it as a variety of that species.

Modern species of *Raphidia* show so much variation in the venation, even on the two sides of the same animal, that fossil

species distinguished by the structure of the wings must be regarded as more or less provisional. Possibly the material will never be sufficient to decide definitely whether we have too many or too few specific names.

Osmylidia requieta (Scudder)

Professor Wickham found a hind wing, 14 mm. long, at the Wilson Ranch.

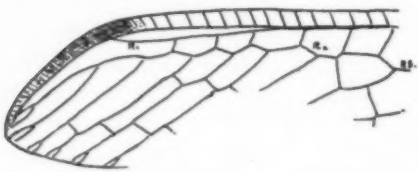

FIG. 1.—*Osmylidia requieta* (Scudd.), hind wing

PALAEOCHRYSA Scudder

The living genus *Allochrysa* Banks cannot be distinguished from this, unless we are prepared to recognize as generic characters which would split the known species into still other genera. *Allochrysa* contains three species: *Palaeochrysa virginica* (Fitch) of the Eastern United States; *P. parvula* (Banks), from Florida; and *P. arizonica* (Banks), from Arizona.

Palaeochrysa fracta n.sp.

Anterior wing about 15 mm. long, 5½ broad; hyaline with dusky stigmatic region; venation pale ferruginous, probably green in life; radial sector originating far basad of "third cubital" cell; costal area broad, terminating about 6 mm. before tip of wing; at least 17 costal cross-veins; about 16 cross-veins between radius and sector, the cells very broad (high), the middle ones twice as high as long; "third cubital" divided by a straight vein in the middle, but the upper division strongly and acutely produced basally, so that the basal angle between the sections is practically a right angle; radial sector giving off 13 oblique branches below; media beginning to zigzag just after leaving "third cubital," its general course gently curved until it meets the radial sector two cells before the apical

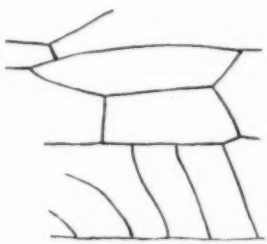

FIG. 2.—*Palaeochrysa fracta*. "Third cubital" cell and adjacent parts.

margin; three veins running from "third cubital" to hind margin; eleven forked veins on hind margin, the first at the tip of the radial sector, at apex of wing; eight simple veins from Cu_1 to hind margin before the first forked one, and one simple vein between the second and third forks.

Miocene shales of Florissant, Wilson Ranch (H. F. Wickham). Nearest to *P. vetuscula* (Scudder), which agrees in having the radial sector originating far basad; but the "third cubital" cell is different.

Palaeochrysa wickhami n.sp.

Anterior wing about 9 mm. long; hyaline, with dark venation; radial sector originating scarcely basad of lower basal corner of "third cubital" cell; costal area with eight cross-veins visible, but at least 10 were present; 9 or 10 cells between radius and sector, the middle ones about square (their shape entirely different from those of *P. fracta*), the basal one long; between R_s and M are seven oblique cells and a basal one of irregular shape (above the "third cubital"); between media and cubitus are nine cells; six simple nervures from cubitus to hind margin before the forked ones begin; "third cubital" divided straight across the middle, neither part much produced basally or apically; one nervure from near middle of "third cubital" to margin, and one meeting the nervure which bounds its outer side. The basal part of cubitus is straight, and the media is not dislocated.

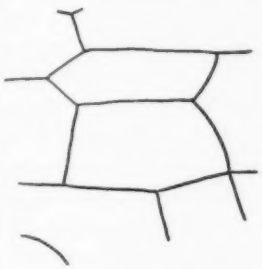

FIG. 3.—*Palaeochrysa wickhami*. "Third cubital" cell.

Miocene shales of Florissant, Wilson Ranch (H. F. Wickham). Remarkable for the small size and dark nervures; it might be a *Nothochrysa*, except for the forked veins going to hind margin. It differs from *P. vetuscula* not only in the much smaller size, but also in not having R_s originating far basad of the double cell, and also in having only one nervure from double cell to margin. In this last character, and also in the shape of lower part of double cell, it differs from *P. ferruginea* and *P. concinnula*.

HYMENOPTERA

Cryptocheilus hypogaeus n.sp.

♀. Robust, about 17 mm. long; anterior wings about 10¼ mm.; antennae thick, about 6½ mm. long, not curled up at end; legs apparently normal for the genus; metathorax with strong transverse grooves. Head and thorax black, abdomen dark, possibly brown in life; antennae with scape dark, flagellum ferruginous; femora dark, tibiae and tarsi probably ferruginous in life; wings with two transverse dark clouds, one at and just beyond the basal nervure, with a lobe extending apicad, the other, large and diffused, below the base of the marginal cell; the large pale area between the bands may have been orange or yellowish, a suggestion of this color remaining; apical region and margin of wing pallid. Venation ordinary; upper angle of second submarginal cell less than a right angle; outer border of third submarginal cell very faint, seen with difficulty. Claws rather short, with a single large sub-basal tooth, the outer side of which presents a gentle slope, instead of being vertical as Bingham figures for *Cryptocheilus nicevillii* (*Salius nicevillii* Bingh.).

The following wing-measurements are in microns. I give also some measurements which I made from the type of *Cryptocheilus laminarum* (*Salius laminarum* Roh.).

Length of marginal cell	2800	(*laminarum* 2640)
Lower side of marginal cell beyond third submarginal	800	(*laminarum* 800)
Greatest width of marginal cell	640	
First submarginal on marginal	224	(*laminarum* 224)
Second submarginal on marginal	800	(*laminarum* 880)
Third submarginal on marginal	1200	(*laminarum* 896)
Third submarginal on third discoidal	640	(*laminarum* 720)
Second submarginal on third discoidal	640	(*laminarum* 690)
Second submarginal on first discoidal	352	(*laminarum* 432)
Lower end of basal nervure basad of transversomedial	720	(*laminarum* 560)
Length of transversomedial	400	
Basal nervure on first submarginal cell	320	
Basal nervure on first discoidal	720	
Greatest (diagonal) length of first submarginal	2896	
Greatest (diagonal) length of first discoidal	2960	

In the hind wing, the approximating ends of the cubital and transversomedial are about 64 microns apart, the cubital more apicad, not more basad as it is in *Cryptocheilus sericosoma* (*Pompilus sericosoma* Sm.).

Miocene shales of Florissant, Wilson Ranch, a very beautiful specimen (H. F. Wickham). In Rohwer's table (*Psyche*, April, 1909) this runs nearest to *Cryptocheilus florissantensis* (*Hemipogonius florissantensis* Ckll.), but differs by the great inequality in the widths of the second and third submarginal cells above, and also by the lack of a dusky cloud on the outer margin of the wing. It is actually nearest to *C. laminarum*, which has a dusky band in the region of the basal nervure, although it is light and inconspicuous in the type. I hesitated at first whether to refer the insect to *C. laminarum*, but there are so many differences in the details of the measurements that I can only regard it as distinct.

It appears that in Miocene times *Cryptocheilus* was producing a number of closely allied species, exactly as in the modern fauna, and in all this time there is no evidence of any generic modification.

Janus disperditus Cockerell

Wilson Ranch (Wickham). Professor Wickham was so fortunate as to pick up the reverse of my original type!

Hemichroa eophila Cockerell

Wilson Ranch (Wickham). About 9 mm. long, anterior wing 7½; thus smaller than the type, but otherwise similar.

HEMIPTERA

Tingis florissantensis n.sp.

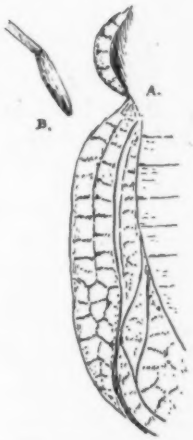

FIG. 4.—*Tingis florissantensis*. A. Side of thorax and elytron; B. end of antenna.

Length of body about 3 mm.; robust, black, of ordinary form; elytra extending perhaps half a mm. beyond abdomen; antennae not especially long, length about 1280 μ, the last joint about 240 μ long; pronotal lateral margins not much expanded,

with a single row of areolae; costal area of elytra with two rows of areolae, except at widest part, beyond middle, where are two sets of three areolae in a transverse series; general form and structure of elytra as in *Gargaphia*.

Miocene shales of Florissant (University of Colorado Expedition). An ordinary looking species, quite distinct from those described by Scudder. Scudder's genus *Eotingis* is singularly like the living *Celantia* Distant, from Ceylon, agreeing in practically everything except the structure of the thorax, which in the fossil genus lacks the vesicular enlargements.

DIPTERA

Protophthiria n.g. (Bombyliidae)

Similar to *Phthiria*, but proboscis stouter, less than twice length of antennae; abdomen much longer than thorax; face longer.

Protophthiria palpalis n.sp.

Length 8 mm., wing 6 mm., robust; head, thorax, and legs dark, probably black in life; shape of head (lateral view) much as in *Phthiria sulphurea*, but face longer (as long as antenna) and occiput more obtuse; antennae much as in *P. sulphurea*, the third joint broad basally and tapering at end, the end narrower than in *P. sulphurea* (antennae therefore quite different from those of *P. pulicaria*); no long hairs at end of antennae; face not hairy; palpi very well developed (990 μ long), formed as in *P. sulphurea*, but proboscis only extending about 480 μ beyond palpi; thorax (in lateral view) not greatly humped, much as in *P. sulphurea*, except that the scutellum is less promi-

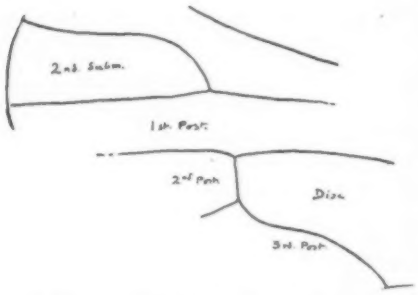

FIG. 5.—*Protophthiria palpalis*, part of apical field of wing.

nent, and there is little or no hair; abdomen thick, about 5 mm. long, with thin but conspicuous pubescence, the hairs quite long, and (as in *P. sulphurea*) especially abundant at end of sixth segment; legs long and slender, hind legs about 8 mm. long.

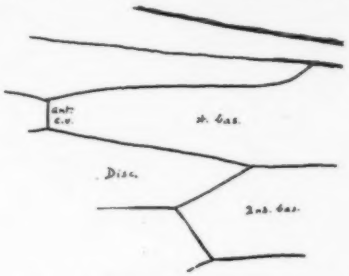

FIG 6.—*Protophthiria palpalis*, ends of basal cells.

Wings hyaline, the costa darkened; two submarginal and four posterior cells; anal cell closed at or almost at margin; second submarginal cell little expanded apically, and not angled above near base; second vein not turned backward at end; first posterior cell widely open.

The following measurements are in microns:

Second basal cell on discal, about	360
Second basal cell on fourth posterior	224
Separation of second and third veins to anterior cross-vein	1,200
Basal corner of discal cell to anterior cross-vein	880
Anterior cross-vein to base of second submarginal cell	1,840
Lower side of second submarginal cell, about	1,680
Transverse (vertical) diameter of second submarginal a little beyond middle	560

Miocene shales of Florissant, Wilson Ranch (Wickham). Among the Florissant fossils this comes nearest to *Lithocosmus*, but the form of the second submarginal and first posterior cells is quite different. The venation is essentially that of *Phthiria* and *Acreotrichus*, while the characters of the head are rather those of *Phthiria* than *Acreotrichus*, though differing from both.

ALOMATIA n.g. (Bombyliidae)

A genus of Lomatiinae with rather long and narrow abdomen; antennae close together, stout, the basal joint very bristly, the apical one elongate-fusiform, bare; wings with two submarginal and four posterior cells; anal cell closed at margin of wing; anterior cross-vein oblique, at about the end of first fourth of discal cell; end of

praefurca considerably (at least 480 μ) basad of anterior cross-vein; second vein bent upward, but its inner angle at end less than a right angle; second submarginal large, including apex of wing; first posterior widely open.

Alomatia fusca n.sp.

Length about 6 mm.; wing $4\frac{3}{5}$ mm.; head apparently pale, no long proboscis visible; thorax and abdomen dark, probably black in

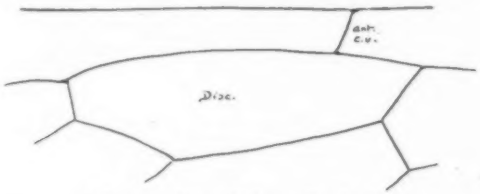

FIG. 7.—*Alomatia fusca*, discal cell

life, abdomen rather elongate, about 4 mm. long, less than 2 mm. broad near base; wings dark fuliginous.

The following measurements are in microns:

Lower side of second submarginal cell	624
Upper end of anterior cross-vein to basal corner of second submarginal cell	1,600
Lower end of anterior cross-vein to apex of discal cell	800
Lower end of anterior cross-vein to base of discal cell	272
Second basal cell on discal	192
Second basal cell on fourth posterior	160

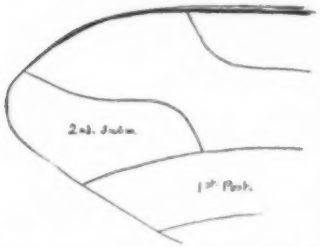

FIG. 8.—*Alomatia fusca*, end of wing

The vein separating the discal cell from the third posterior shows a single uniform gentle curve.

Miocene shales of Florissant, Wilson Ranch (Wickham). In my manuscript table of Bombyliid genera (based on venation) this runs to the vicinity of *Oncodocera*, *Lithocosmus*, and *Protophthiria*. It is, however, remarkable for the position of the anterior cross-vein. From *Protophthiria* it is also easily

known by the different termination of the second vein, and many other details. From *Lithocosmus* it is known especially by the first posterior cell being without apical contraction. In Verrall's table of Lomatiinae it runs closest to *Prorachthes*. The praefurca of *Alomatia* ends at least 240 μ basad of basal corner of discal cell, which is contrary to Verrall's diagnosis of Lomatiinae, but his own figure of *Lomatia lateralis* shows a similar condition. The antennae of *Alomatia* are curiously like those of *Thereva*.

FIG. 9.—*Alomatia fusca*, antenna.

PROTOLOMATIA n.g. (Bombyliidae)

A genus of Lomatiinae, related to *Lomatia*, to which it runs in Verrall's table of Lomatiinae. It differs in the long, parallel-sided abdomen, and the much less oblique anterior cross-vein. Among the fossil genera it is nearest to *Megacosmus*, the species of which are much larger, with broader abdomen, and have the end of the second vein curved strongly backward (not far from vertical in the new genus), and the side of the second basal cell on discal very long.

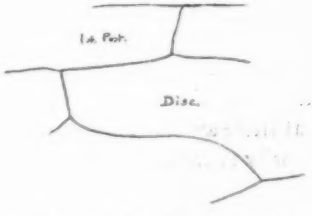

FIG. 10.—*Protolomatia antiqua*, apex of discal cell.

Protolomatia antiqua n.sp.

Length about 9 mm.; wing $6\frac{3}{4}$ mm.; body dark, slender; abdomen parallel-sided, its width $1\frac{3}{4}$ mm.; antennae slender, about 1200 μ long; thorax and abdomen without evident hair. Wings hyaline, faintly dusky; venation nearly as in *Lomatia lateralis*, but praefurca longer, anterior cross-vein much less oblique (about as in *Megacosmus secundus*), upper apical side of second basal cell shorter.

The following measurements are in microns:

Length of praefurca	512
First posterior cell on first submarginal	1,250
First posterior cell on second submarginal	1,150

Width of first posterior cell on outer margin of wing	288
Discal cell on first basal	1,360
Discal cell on first posterior	592
Second basal on discal	240
Second basal on fourth posterior	160
Anal cell on wing margin	400

Miocene shales of Florissant, Wilson Ranch (Wickham). The remarkably numerous and diversified Florissant Bombyliidae still fail to include any Anthracinae. Probably, as parasites, these Bombyliids took the place of the Tachinidae, which were rare if not wholly absent.

FIG. 11.—*Protolomatia antiqua*, apex of wing.

Asilus peritulus Cockerell

Wilson Ranch (Wickham). A wing, about 11½ mm. long.

Hirmoneura melanderi Cockerell

Wilson Ranch (Wickham). Total length 19 mm.; the abdomen long and tapering.

Syrphus willistoni Cockerell

Wilson Ranch (Wickham). Length about 10½ mm.; wing a little over 8½. Other specimens which I have referred to this species are considerably smaller; I believe that there are two species of this type, but at present I cannot satisfactorily separate them.

Petrological Abstracts and Reviews

Edited by ALBERT JOHANNSEN

Winchell, Alexander N. *Directions for Laboratory Work in Optical Mineralogy.* Published by the Author, Madison, Wis., 1911. Pp. 36. Wrappers.

This very useful little pamphlet for laboratory use was first published for students at the University of Wisconsin. It is now reprinted with the omission of the original slide numbers but with blank spaces into which may be written those of any collection of thin sections. Problems are given in the use of the microscope for determinations by ordinary light, parallel polarized light, crossed nicols with parallel light, and crossed nicols with convergent light. Throughout the book references to the proper pages in the author's *Optical Mineralogy* are given for the explanation of the phenomena observed.

A. J.

Winchell, Alexander N. *Geology of the National Mining District, Nevada.* Min. and Sci. Press, 1912. Pp. 16.

Describes various rocks occurring in Nevada. Suggests the name auganite for "a volcanic rock consisting essentially of basic plagioclase and pyroxene. The plagioclase is usually labradorite, but may be more basic. . . . . Having provided the new name for rocks ordinarily called augite andesite, it is possible to use the latter name for rocks which are actually varieties of andesite, and consist essentially of acid plagioclase and pyroxene." The reviewer believes that augite andesite, in the past, has been so used by most petrographers. Auganite, therefore, appears to be suggested for olivine-free basalt, the dividing line between andesite and basalt being plagioclase of the composition Ab_1An_1.

A. J.

Woyno, Tadeusz Jerzy. "Petrographische Untersuchung der Casannaschieter des mittleren Bagnetales (Wallis)," *Neues Jahrb.*, XXXIII (1911), 136–207.

The Casanna schists in the central Bagnetals consist of a complex of metamorphosed rocks whose chief constituents are glaucophane, epidote,

chlorite, albite, sericite, quartz, and calcite. In most of the rocks the amount of soda is greater than potash. Of these rocks, two, prasinite and epidote glaucophane schist, are derived from igneous rocks.

A. J.

WRIGHT, FRED. EUGENE. "Mikroskopische Petrographie vom quantitativen Gesichtspunkte aus," *Neues Jahrb.*, XXV (1913), 753–75.

A translation of an article by the same writer in the *Journal of Geology*, XX (1912), 481–501.

WRIGHT, FRED. EUGENE. "A New Thermal Microscope for the Measurement of the Optical Constants of Minerals at High Temperatures," *Jour. Wash. Acad. Sci.*, III (1913), 232–36.

Describes and illustrates a new thermal microscope for the determination of the birefringences, extinction angles, and optic axial angles of minerals at temperatures between 10° C. and 1200° C.

A. J.

WRIGHT, FRED. E., and VAN ORSTRAND, C. E. "The Determination of the Order of Agreement between Observation and Theory in Mineral Analyses," *Jour. Wash. Acad. Sci.*, III (1913), 223–31.

The authors discuss the principles underlying the calculation of mineral formulas.

A. J.

WRIGHT, FRED. EUGENE. *The Methods of Petrographic-Microscopic Research*. Carnegie Institution, Publication No. 158. Washington, 1911. Pp. 204, pls. 11.

The author found, in his work on silicate preparations in the Geophysical Laboratory, that the ordinary methods of determining minerals microscopically were not always sufficient in working with artificial fine-grained products, and it was necessary to devise new methods or modify old ones. While many of the results of the author's comparative studies on the relative merits and accuracy of a number of these methods have been published at different times in the *American Journal of Science*, they are here brought together in proper relations to other available methods. As stated by the author, the different methods best adapted

for the microscopic examination of fine-grained and artificial preparations are here considered with special reference to their degree of accuracy and range of general application.

In the introduction, the petrographic microscope and its several mechanical and optical parts are described. This is followed by chapters dealing with the methods for determining refractive indices, birefringence, and extinction and optic angles. There are also reproduced some extremely useful charts, such as Wulff's stereographic net, Hilton's gnomonic projection net, Fedorow's refractive index diagram, and various others.

The book is an extremely valuable contribution to the literature of petrography, and is especially important for the author's comparative study of the value of different methods. It may be studied with profit by all advanced students.

A. J.

WRIGHT, FRED. EUGENE. "The Index Ellipsoid (Optical Indicatrix) in Petrographic Microscopic Work," *Amer. Jour. Sci.*, XXXV (1913), 133–38.

The writer suggests abandoning the "elasticity ellipsoid" and the symbols for the "axes of elasticity" in the explanation of the phenomena of light in crystals. He would use instead only the indicatrix and the symbols for the refractive indices, regarding the use of other symbols as bewildering to the student. The reviewer's experience has been that students can grasp much more readily the idea of an ease (or difficulty) of vibration in a certain direction in a crystal, and a corresponding rate of movement at right angles to it, than they can the inverse relation of the refractive indices. The reviewer long ago abandoned the terms "axes of elasticity" and substituted for them "ease of vibration axes."

A. J.

WRIGHT, FRED. EUGENE. "Oblique Illumination in Petrographic Microscope Work," *Amer. Jour. Sci.*, XXXV (1913), 63–82.

A study of the cause and effect of oblique illumination in the study of mineral plates. A sliding diaphragm in the focal plane of the condenser is considered by the author to be the best method for producing this illumination. He points out that in the measurement of extinction angles, central illumination with parallel plane-polarized light is essential.

A. J.

WÜLFING, E. A. "Fortschritte auf dem Gebiete der Instrumentenkunde," *Fortschritte der Min., Krist., u. Petrog.*, III (1913), pp. 63–92.

A summary of the progress made, during the past decade, in mineralogical and petrographical instruments, including goniometers, microscopes, microscope accessories, conoscopes, total reflectometers, photometers, chromoscopes, lighting-, projection-, grinding-, and separating-apparatus, sclerometers, electromagnets, electric ovens, drawing instruments, etc. A bibliography of 214 articles, and brief descriptions of the various instruments are given.

A. J.

WÜLFING, E. A. "Eine einfache Vorrichtung für konstante Wasserbäder," *Zeitschr. f. angew. Chemie*, XXVI (1913), pp. 87–90.

Describes a new water bath for constant temperatures.

A. J.

WÜLFING, E. A. "Ueber die objektive Darstellung der Grenzkurven bei Kristallen," *Sitzb. Akad. Wiss. Heidelberg*, 1912, Ab. 19, pp. 14.

WÜLFING, E. A. "Ueber Projektion mikroskopischer Objekte ins besondere im polarisierten Licht," *Sitzb. Akad. Wiss. Heidelberg*, 1911, Ab. 36, pp. 40.

Describes a lantern for projecting microscopic objects. With this instrument it is possible to project interference colors of minerals, interference figures, to show the optical character of minerals, and in fact to show upon the screen all of the phenomena seen with the polarizing microscope.

A. J.

WÜLFING, E. A. "Ueber kristallographische Kaleidoskope," *Neues Jahrb.*, I (1912), 37–50.

Describes an apparatus by means of which crystal form, crystal symmetry, etc., may be clearly shown to students.

A. J.